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Carbon Dioxide Research Division

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024

Workshop on Sea Level Rise and Coastal Processes

Palm Coast, Florida
March 9-11, 1988

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SUMMARY

The possibility in the coming decades of a higher rate of relative sea-level rise globally is now thought to be sufficiently great to warrant serious consideration for its potential implications to civilization. With regard to shoreline response to relative rise as well as the rate of rise, questions emerge almost immediately about how the open coast and estuarine shorelines would change. Since, for example, a significant portion of the U.S. coastline is composed of loose materials including sand and muddy sediment, it is evident that simple inundation models based on existing terrestrial topography would be far from adequate in predicting shoreline configurations for any given sea-level rise scenario. Consequently, it becomes essential to examine the state-of-the-art technology in shoreline prediction modeling, gaps in fundamental knowledge of coastal processes, and future research needs for advancing the technology to accomplish the task of prediction to a meaningful level of accuracy.

To answer these queries the University of Florida (UF) undertook a study during the summer of 1987 which resulted in a report, "Some Considerations on Coastal Processes Relevant to Sea Level Rise," authored by Ashish Mehta, Robert Dean, William Dally, and Clay Montague. In this report the effects of potential sea-level rise on the shoreline and shore environment were briefly examined by considering the interactions between sea-level rise and coastal processes. These interactions were reviewed, beginning with a discussion of the need to reanalyze previous estimates of eustatic sea-level rise and compaction effects in water level measurement. This was followed by considerations on coastal and estuarine tidal ranges, storm surge and water level response, and marine interaction with natural and constructed coastal features. The desirability to reexamine the well-known Bruun rule for calculating shoreline recession due to the likelihood of significant cross-shelf sediment transport was recognized. The mechanics of salt penetration in groundwater and surface water was reviewed, followed by effects of sedimentary processes in the estuaries including wetland response, particularly in the fine-grained environment. Finally, comments were included on some probable effects of sea-level rise

on coastal ecosystems, since response in this case is unquestionably contingent upon hydrodynamic and sedimentary forcing.

These considerations amply demonstrated the complexities of the interaction between sea-level change and loose boundary shoreline, their site-specificity, and the inadequacy of inundation models. It was also concluded that, with some minor exceptions, the basic knowledge of coastal processes and the available data base, including hydrodynamics, sedimentary processes, and their interaction, are inadequate for predictive modeling. Apart from difficulties in modeling boundary layer turbulence and associated mixing processes, sediment transport formulations require knowledge of a host of free coefficients which tend to be highly site-specific and therefore difficult to evaluate in the complex coastal environment. As a case in point, our ability to predict long-term shoreline evolution is hampered by the accuracy with which local littoral drift distribution can be predicted, despite great strides made in this area of research in the past couple of decades. Besides further improvements in theory, the need for a better definition of wave forcing through adequate long-term monitoring of the coastal wave field via field measurements cannot be overemphasized.

To critique the conclusions of the UF report and to reach a broader consensus on research needs to vastly improve shoreline response predictive capabilities, a Workshop on Sea Level Rise and Coastal Processes was held at Palm Coast, Florida, on March 9-11, 1988. After the workshop the UF report was modified by the authors to constitute the bulk of the present report. An appendix to the report contains comments by the workshop participants, excluding the authors of the UF report. These comments comprise reviews of the UF document and views of individual participants regarding specific subject areas of their expertise, including research needs. While all the participants were chosen for their international experience and awareness of the problem, specific mention must be made of the contribution of John de Ronde, who was invited to present his view of the Dutch experience, which in many ways should provide insight into what may occur in other parts of the world at higher relative sea level.

Participant comments largely reinforce the conclusions of the UF contribution through additional illustrative examples, clarifications, and

qualifications in some cases. The role of possible climatic changes in addition to sea-level rise in influencing shoreline episodic response has been pointed out by several participants. Some have also touched upon the need to take sea-level rise into account in designing coastal structures over a typical design lifespan of 50 years. This issue has been examined in some detail in a recent (1987) report of the National Research Council Marine Board, "Responding to Changes in Sea Level: Engineering Implications," which may be consulted for that purpose.

As pointed out during the workshop, unlike the situation in the United States, a significant fraction of the global shorelines is not composed of loose material; beaches are seemingly more important, economically and otherwise, in some countries than others. Required spatial scales for long-term shoreline response prediction vary greatly with human and ecosystem needs. Nevertheless, it can be undoubtedly concluded that prediction technology for this purpose is nascent and awaits further research emphasis.

1. INTRODUCTION

The complexities of shoreline response to sea level rise are contingent upon a very wide range of inter-relationships between physical/ecological factors. The focus of resource analysis for the present purpose must ultimately be on predictive capability, since we are principally dealing with the question of how shorelines and shore environment will change with future sea level rise. Prediction in turn requires an understanding of process fundamentals and adequate data. Therefore, much of what follows pertains to these aspects, which in many cases have more to do with the basics of resource response to hydrodynamic and meteorologic forcing than to sea level rise. If this can be elucidated, then imposing and evaluating the effect of sea level rise becomes a far less difficult task.

Organization of basic knowledge is intertwined with the question of resolution of spatial and temporal scales. The desired resolution for the evaluation of a resource is set by criteria which are dependent upon many non-technical factors. At a built-up shoreline, a 10 m recession could severely damage a structure, while at a natural shoreline the concerns will be less stringent. Then again, in low lying areas such as the Florida Everglades, just a few centimeter rise in sea level would prove to be disastrous to water management, and would cause extensive ecological changes associated with salinity intrusion. A rapidly rising sea level can generate a materially different response than a slow one, an example being the fragile barrier island shoreline. Finally, there is the question of absolute sea level rise and the associated shoreline scenarios. By keeping the issues focussed on the coastal processes themselves, we have in the most part stayed clear of centering on specific temporal and spatial scales explicitly, even though such considerations are inherent in evaluating the degree of uncertainty in the state-of-the-art knowledge and in future research needs.

The interactive nature of coastal processes renders it difficult to isolate resource issues and place them under well-defined "umbrellas" for descriptive purposes. We have selected ten headings (sections 2 through 11) within which a range of topics has been referenced. The first of these

- Estimates of Eustatic Sea Level Rise - does not deal with process description in a general way, but highlights a fundamental issue, namely the quality of the data base that has been used to calculate past secular trends in sea level change, and what needs to be done to improve this base. Following this is the section Compaction Effects, which is directly associated with problems in water level measurement.

Sections 4 though 11 deal with coastal processes. In section 4 the effect of sea level rise on tidal ranges is discussed, and section 5 deals with non-astronomical factors including storm surge and waves. The next two sections are concerned with shoreline response. While section 6 deals with physical processes in shoreline response in broad categories, section 7 focusses on specific issues relative to the scope and limitations of the well known Bruun Rule for estimating shoreline recession rate. Physical considerations upon which this rule must be re-examined have been noted.

Section 8 describes problems with saltwater intrusion in groundwater as a result of sea level rise or analogous effects, while the same problem in surface waters is highlighted in section 9. Sedimentation problems in tidal entrances, estuarine mixing zone and wetlands is described in section 10. Finally, ecological changes, including research needed to quantify these better, have been noted in section 11.

Some overlap between the various sections is inevitable. This extends to both the physical description and research needs. Also, by and large, the coastal processes have been reviewed from an engineering perspective, and evaluation of present day knowledge has been made from the viewpoint of the availability of quantitative (as opposed to qualitative) criteria.

In general it appears that with the possible exception of tidal hydrodynamics and salinity intrusion, considerable further research is required for assessing shoreline and shore environmental response in a confident manner. Strides made during the past decade have been impressive, but for example where sediment transport is a key factor, we are significantly limited in long-term predictive capability. This is partly due to the lack of good quality synoptic hydrodynamic/meteorologic data. This problem in turn has an impact on ecological modeling, which is contingent upon a knowledge of flows and sediment movement.

The bibliography is divided by sections. In some cases, additional references not cited in the text, but considered to be of potential interest to the reader, have been included.

In order to reach a broader consensus among scientists concerned with the subject area, a Workshop on Sea Level Rise and Coastal Processes was held at Palm Coast, Florida, on March 9-11, 1988. The main objective was to use the findings of the 1987 University of Florida report, "Some Considerations on Coastal Processes Relevant to Sea Level Rise," authored by Ashish Mehta, Robert Den, William Dally, and Clay Montague, as a basis for a wide ranging discussion on relevant available data, their quality, predictive state-of-the-art and future research needs. Several of the participants were asked to provide written inputs to the different chapters (2 through 11). Their reviews and related comments have been appended. Comments/discussions were provided by David G. Aubrey, Robert Biggs (who was unable to attend the meeting), Robert A. Dalrymple, Vivien M. Gornitz, William McAnally, Louis H. Motz, Thomas J. Smith (co-authored by W. M. Kitchens, who did not attend) and Jacobus van de Kreeke. Finally, a summary of the presentation by John G. de Ronde from the Rijkswaterstaat, the Netherlands, is included. In all cases these comments/discussions supplement material presented in the chapters by way of critique, new insights and case studies.

2. ESTIMATES OF EUSTATIC SEA LEVEL RISE

2.1 INTRODUCTION

Eustatic sea level rise is the global average sea level rise primarily due to: 1) additional water mass in the oceans through release of water contained in polar ice caps and alpine glaciers, and 2) steric expansion of water presently in the oceans due to increased temperature, thereby increasing the volume of an existing water mass. Sea level change data from 20,000 years before present (BP) to 1,000 years BP have been obtained from radiometric dating of plants and animals that lived only in intertidal or shallow marine waters. Data from the last 100 or so years are based on measurements from long-term tide gages. Both of these sources include not only the "signal" of eustatic sea level change, but the "noise" or contamination by local vertical movement of the land where the measurements are made. Additionally, local and temporal oceanographic and meteorological factors may contribute to anomalously high or low water levels for periods of many years. The degree of contamination in any one tide gage record may be severe with the annual contamination exceeding up to 40 years of eustatic trend. Much of the contamination is spatially and temporally coherent over fairly long distance and time scales and the physics of this contamination is poorly understood. If the available tide gage data provided a representative distribution over the world's oceans, the noise could be eliminated by simply averaging over these gages. However, the available tide gage data are heavily concentrated in the northern hemisphere and along continental margins.

Tide gages measure the local relative sea level which is important and is the water level relevant to that area. However, an understanding of recent eustatic sea level rise is critical, because models developed for predicting future sea level rise are calibrated based on estimates of recent rise. Most of these estimates suggest a rate of 10-15 cm/century (1 to 1.5 mm/yr) with some investigators inferring an increase in the rate of rise over the past 40 or so years. Most of the studies leading to the above estimates have been based on gages located in reasonably stable low- to mid-latitude areas. Clearly the most significant neotectonic

contribution to relative sea level rise is the earth's rebound from the ice loading in the polar regions during the last (Wisconsin) ice age. This rebound is causing uplift in the high latitudes on the order of 1 meter per century and land subsidence at the lower latitudes on the order of 5 cm per century. There have been suggestions that most of the studies of eustatic rates, in excluding the high latitudes of relatively rapid uplift, have yielded overestimates. A very preliminary analysis presented here based on United States data tends to support this contention.

Areas in which future studies appear warranted include: 1) understanding the physics of the noise in tide gage records with the objective of extracting this portion of the record, 2) revisiting the question of extracting recent eustatic sea level rise rates from the tide gage records with an emphasis on proper recognition of the contribution from glacial rebound at all latitudes, and 3) if the changes resulting from 2 are significant, recalibrating the models employed for predicting future sea level rise based on scenarios of future changes in CO_2, other trace gases and a gradual warming trend.

2.2 LITERATURE REVIEW

There has been a wide range of techniques and degree of sophistication applied in an attempt to extract eustatic sea level rise (ESLR) from tide gage records. One of the first comprehensive published studies on ESLR based on tide gages was by Gutenberg (1941). A total of 69 gages was analyzed encompassing the period 1807 to 1937. Gutenberg excluded tide gages known to be in areas of crustal uplift, yet gages were included in areas known to be sinking, some at fairly high rates. Gutenberg concluded that ESLR was approximately 1 mm per year.

Many investigations following those of Gutenberg have tended to adopt his data selection procedures with similar results, i.e., rates of 1 to 1.5 mm/yr, see Table 2.1. Emery (1980) concluded that ESLR has been accelerating with a rate up to 3 mm/yr over the past 40 years. Subsequent studies by Aubrey and Emery (1983) and Barnett (1983) conducted specifically to examine the change in rate concluded there was no convincing evidence for such a conclusion.

Table 2.1. Estimates of Eustatic Sea Level Rise Based on Tide Gage Data (adapted from Barnett, 1983; and Hicks, 1978)

Author(s)	Estimate (cm/100 yr)
Thorarinsson (1940)	> 5
Gutenberg (1941)	11 ± 8
Kuenen (1950)	12 to 14
Lisitzin (1958)	11.2 ± 3.6
Fairbridge and Krebs (1962)	12
Hicks (1978)	15 (U.S. only)
Emery (1980)	30
Gornitz et al. (1982)	12 (10 cm excluding long-term trend)
Barnett (1983)	15

The difficulties of extracting the sea level rise (SLR) "signal" from a record containing substantial noise has been studied carefully by Sturges (1987). The coherency of spatially separated tide gage records was investigated with the hypothesis that coherent signals with no lag could be interpreted as global sea level rise whereas lags with a certain character could be interpreted as due to atmospheric forcing or long water wave (Rossby wave) motions. As an example, the records at San Francisco and Honolulu were found to be coherent at periods of 5 to 10 years and longer, although with a phase lag. A comparison of the energy spectra obtained from these two stations is presented as Fig. 2.1a and other spectral information is presented in Figs. 2.1b,c,d. The amplitudes of these coherent components are 5-15 cm. Similar coherence results were found for tide gage records located on both sides of the Atlantic. Sturges concluded that the available records are contaminated by substantial energy with periods up to 40 to 50 years, thus exacerbating the problem of identifying any change in the rate of SLR. The ability to extract the SLR signal may possibly be enhanced through an analysis which recognizes the probable cause of the noise components, thereby guiding their removal from the record.

Aubrey and Emery (1983) applied the method of eigenanalysis to United States tide gage data in an attempt to identify fluctuations that were spatially and temporally coherent. This method, among the most sophisticated applied to date, has the potential advantage of retaining in the first few temporal eigenfunctions, those fluctuations that have the same form and that are either exactly in or exactly out of phase. The principal disadvantage is that the method is purely statistical and does not recognize the physics of the phenomenon, although it may isolate features that will assist in identifying physical components. A particular drawback is that the method only recognizes correlations which are either in phase or exactly out of phase as "signal." Thus a very long and slowly propagating wave would be rejected as noise whereas a pure standing wave would be recognized as "signal." Aubrey and Emery first applied the technique to 12 U.S. gages each of which encompassed 61 years of data and secondly to 41 tide gages with a common time base of 40 years of data. Different rates of rise were found for the East and West coasts. From the

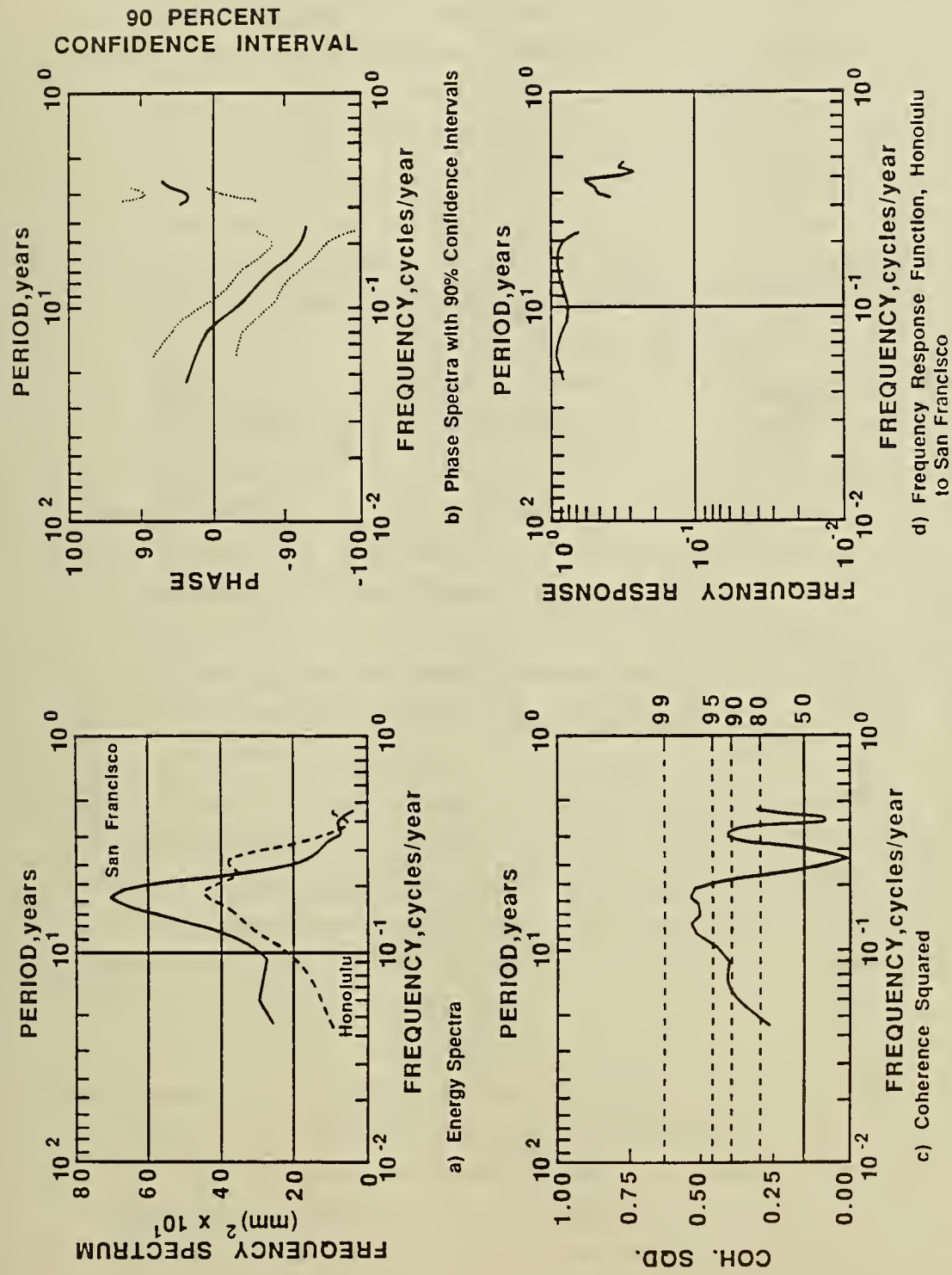


Fig. 2.1. Cross-Spectral Characteristics Between Sea Level at San Francisco and Honolulu: Yearly Data, 1905 Through 1971 at San Francisco and Beginning 1907 at Honolulu (reprinted with permission from Sturges, 1987, "Large-scale coherence of sea level at very low frequencies," Journal of Physical Oceanography, Volume 17, copyright American Meteorological Society).

longer term data set of 12 stations, the eustatic values on the West and East coasts were found to be rising by averages of 1.4 mm/year and 1.3 mm/year, respectively. For the shorter term (40 years) of 41 stations, the rates of change for West and East coasts were -0.3 mm/yr and +2.5 mm/yr, respectively. It was found that the long-term rates of sea level rise are increasing from Cedar Key on the Florida west coast to Cape Hatteras, decreasing from Cape Hatteras to Cape Cod and increasing from Cape Cod to Eastport, Maine. These results are presented in Fig. 2.2. Finally, it was concluded that there is no evidence from this analysis that rates of SLR are increasing over the past 10 years.

Pirazzoli (1986) has analyzed the results from 1,178 tide gage stations provided primarily by the Permanent Service for Mean Sea Level. This appears to be the largest data set considered in an individual analysis. The analysis method was straightforward, first taking averages for each station over five year periods, then averaging over the two ends of the resulting data to obtain a change in sea level from which the rate is determined. The results are presented regionally and on a global basis. The effects of glacio-eustatic adjustment to the last ice age are very apparent in the data with relative sea level (RSL) rising and lowering in most low and high latitudes, respectively. The possible effects of earthquakes in causing sudden displacements and altering the trend after the earthquake are illustrated. As an example, the tide gage at Messina, Italy recorded an abrupt increase in RSL of 57 cm during the earthquake of 1908. Anthropogenic effects, primarily the extraction of water and hydrocarbons, causing compaction are noted with Venice, Italy particularly evident as a consequence of ground water pumping. In attempting to infer global rates from the available data, it is noted that if the earth is divided into 30° latitude and longitude sectors, a total of 72 compartments result of which 71 have marine coasts. The data distribution in these compartments is very non-uniform. Most of the tide gages (70%) are situated in only 4 compartments whereas there are no data in 70% of the compartments. Long-term tide gage data in the southern hemisphere are particularly sparse with over 97% of the stations examined by Pirazzoli in the northern hemisphere. Without the assumption that the results from the northern hemisphere are globally representative, the available data are

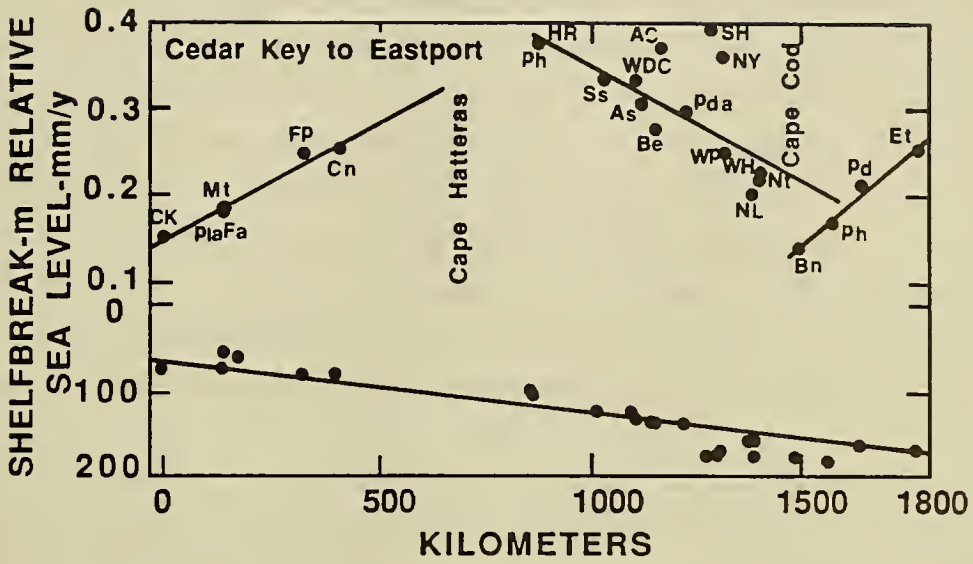


Fig. 2.2. Mean Annual Relative Sea Level Changes During 40-Year Record. Lines Define Three Main Segments of East Coast with Differing Sea Level Trends. Reprinted with permission from Continental Shelf Research, Volume 2, D. C. Aubrey and K. O. Emery, "Eigenanalysis of recent United States sea levels," copyright 1983, Pergamon Press plc.

clearly inadequate. Fig. 2.3 presents a distribution of the tide gage locations according to the longitude-latitude compartments noted earlier. Fig. 2.4, also from Pirazzoli, presents the distribution of tide gages and median trend of RSL by 5° increments of latitude. The earlier noted effect of relative rises in the mid-latitudes and lowering RSL in the higher latitudes is evident.

Pirazzoli concludes that the results presented by most investigators ($\gg$1 mm/yr) probably are an overestimation of the ESLR. Local and regional factors including tectonic movements and oceanic factors are generally larger than eustatic factors. The bias due to downwarping as a result of loading of the continental shelves by sediment transport and deposition is noted. Finally, when centimeter accuracy is attainable from satellite altimetry, the potential to contour the open ocean is regarded as a major advance in our general knowledge of eustatic sea level rise rates which have both good geographic coverage and are free from much of the contamination which attends measurements of tide gages located along the coastline.

Lambeck and Nakiboglu (1984) have carried out an analysis of the effect of post-glacial adjustment on estimates of ESLR. For this purpose, a viscous model of the earth was adopted with the assumption of a uniform mantle viscosity. To quantify the effect of rebound on estimates of ELSR as determined from tide gage records, the apparent or RSL rises predicted by the model without any additional water mass or steric changes were computed for the same eight long-term tide gage stations selected by Barnett (1983). Two values of viscosity, μ, were used: Model 1, $\mu = 5 \times 10^{21}$p and Model 2, $\mu = 10^{22}$p. For the eight stations, Models 1 and 2 predicted apparent (relative) sea level rises of 0.5 and 0.8 mm/yr, respectively whereas Barnett found 1.5 mm/yr. Based on this comparison, Lambeck and Nakiboglu conclude that the post-glacial rebound contribution may be as high as 30% to 50% of published estimates of ESLR.

A limited analysis has been carried out here to attempt to determine the effects of employing only the lower latitude tide gate data. The U.S. data for the East and West coasts and Gulf of Mexico as published by Hicks et al. (1983) were used. The trend estimates in Hicks et al. were

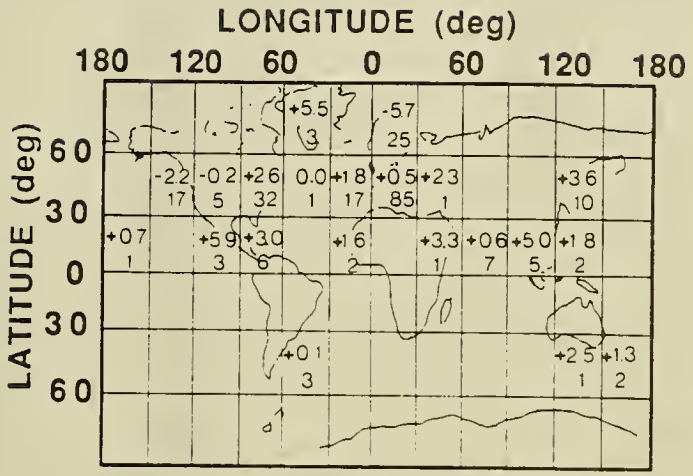


Fig. 2.3. Characteristics of Tide Gage Data by 30° Longitude and Latitude Sectors. The Lower Values Represent the Number of Tide Gages in Each Sector. The Upper (Signed) Numbers Represent the Linear Long-Term Relative Sea Level Change Resulting from Those Gages (reprinted with permission from Pirazzoli, 1986).

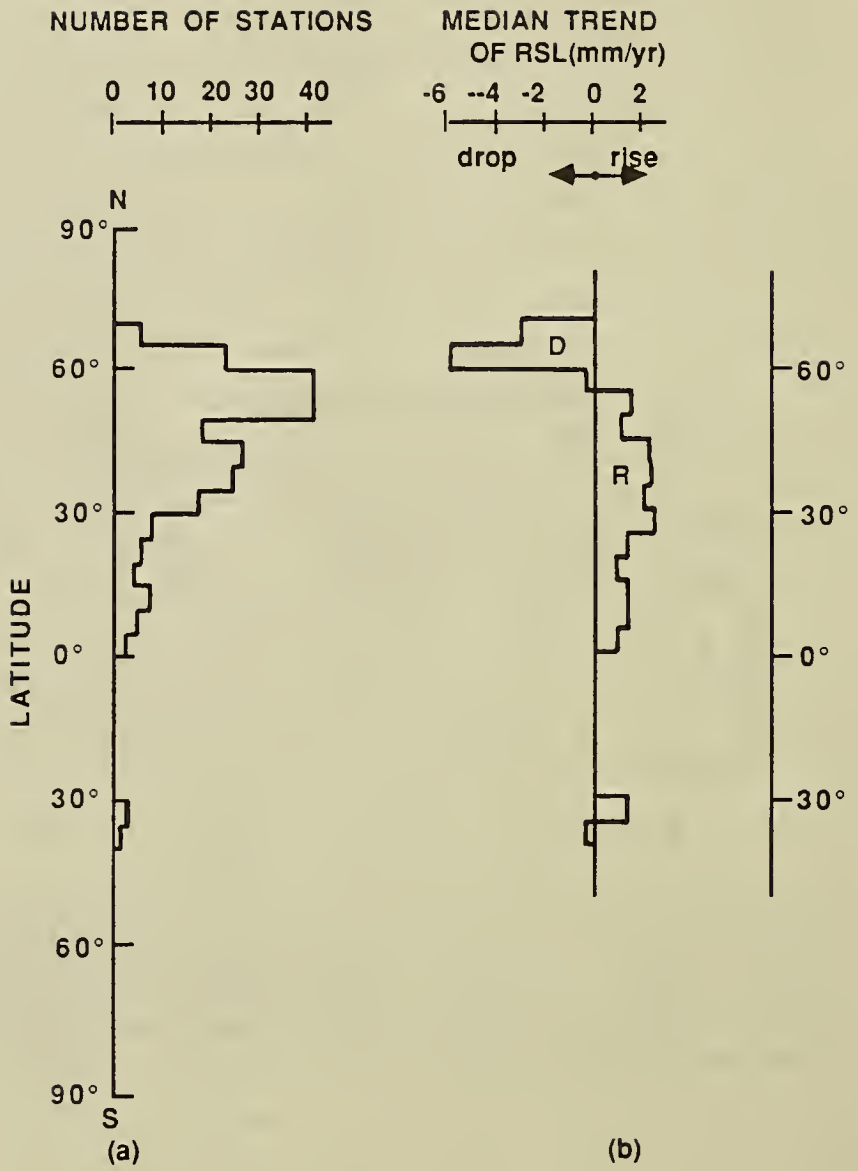


Fig. 2.4. Distribution by 5° Latitude Belts of a) Tide Gage Stations, and b) Median Values of Linear Long-Term Trends of Relative Sea Level. Note the Tendency for a Relative Drop in Sea Level for the Higher Latitudes (reprinted with permission from Parazzoli, 1986).

simply plotted against latitude as presented in Fig. 2.5. A problem is that the data only encompass latitudes from approximately 25° to 58° and thus it is necessary to extrapolate liberally. At the lower latitudes, the data were extrapolated uniformly at approximately 3.2 mm/yr and at the higher latitudes, due to the uncertainties, two extrapolations were adopted to determine sensitivity as presented in Fig. 2.5. Based on the latitudinal variation, $\dot{\eta}(\phi)$, estimates of the ESLR, $\dot{\eta}_E$, were based on the following

$$\dot{\eta}_{E_j} \approx \int_0^{\pi/2} \dot{\eta}_j (\phi) \cos\phi \, d\phi \tag{2.1}$$

where j = I,II represents the different high latitude extrapolations. The resulting values were

$$\dot{\eta}_{E_I} = 0.32 \text{ mm/yr, Extrapolation I}$$

$$\dot{\eta}_{E_{II}} = 0.67 \text{ mm/yr, Extrapolation II}$$

These results are qualitatively in agreement with those of Lambeck and Nakiboglu.

2.3 THE NATURE AND ANALYSIS OF SEA LEVEL DATA

From the standpoint of extracting eustatic sea level change, it is useful to represent the total RSL, $\eta_i(t)$, as measured by the i^{th} tide gage as

$$\eta_i(t) = \eta_E(t) + \eta_{N_i}(t) \tag{2.2}$$

in which $\eta_E(t)$ is the eustatic sea level at time t and $\eta_{N_i}(t)$ is the total "noise" at the i^{th} tide gage. The noise can contain many components including vertical ground motion, effects of freshwater in the vicinity of the gage, coastal currents, long waves, barometric pressure anomalies, wave effects, etc. Several obvious results follow from Eq. 2.2. First, if there were a uniform coverage of tide gages on the oceans, an average of the elevations from all such tide gages would yield the eustatic sea level. Additionally, the eustatic sea level change rate need not be constant, but

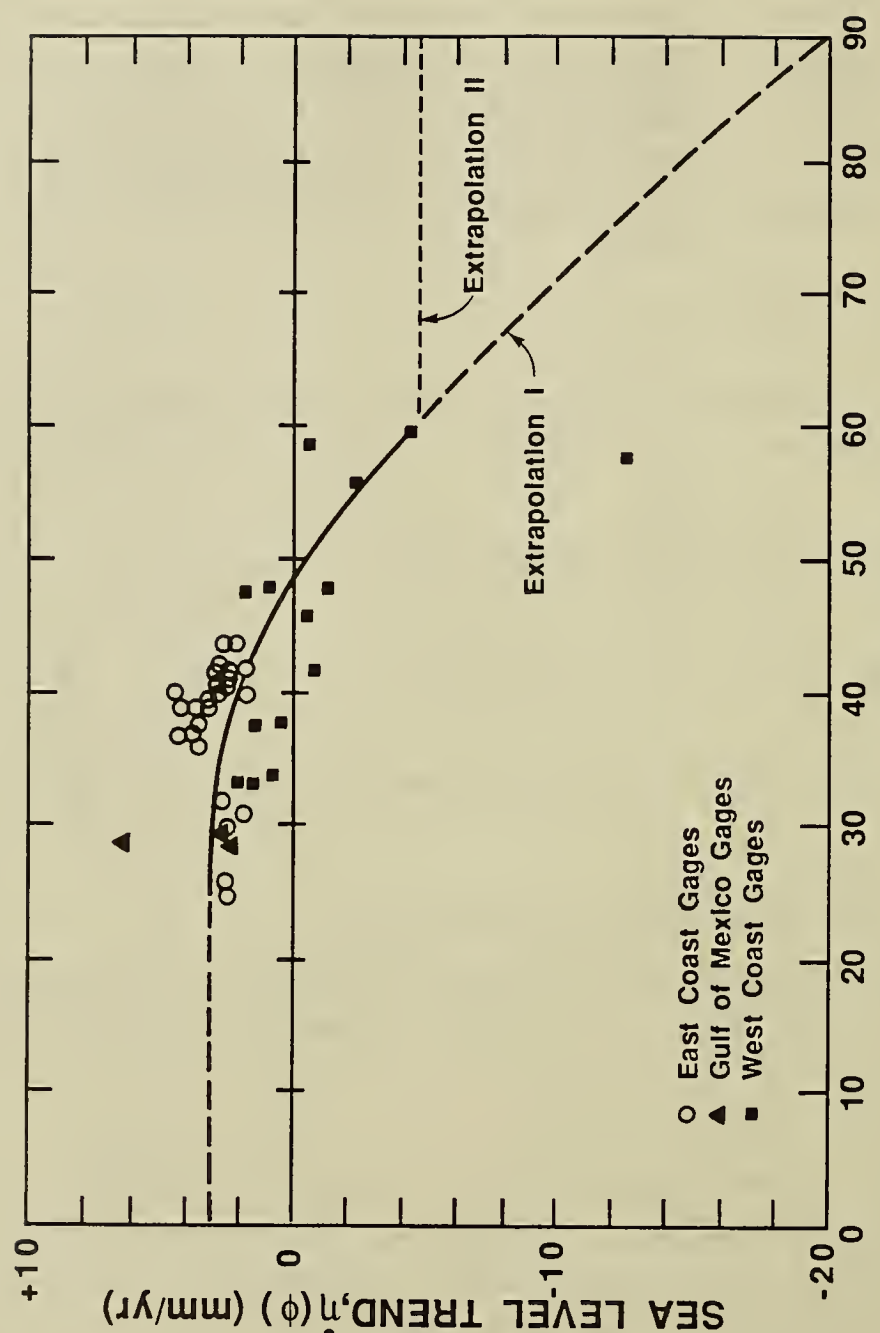


Fig. 2.5. Long-Term Tide Gage Trend Results, $\dot{\eta}$, versus Latitude, ϕ. Continental United States and Alaska (based on Hicks et al., 1983).

could vary substantially year-to-year with temperature, etc. Considering two or more tide gages, the noise may be correlated in space and time positively, negatively, with an arbitrary phase or uncorrelated. The more widely separated the gages, the greater the likelihood that the noise will be uncorrelated. Thus, there are advantages to averaging many records along a coast, possibly with an appropriate coastal length weighting factor. Finally, the best estimate of eustatic sea level (and thus eustatic sea level rise) and one which yields the most understanding as to the stability of the results is a progressive averaging in which larger and larger data bases are averaged, i.e.,

$$\bar{\eta}_{IK}(t) = \frac{\sum_{i=1}^{IK} \eta_i(t) w_i}{\sum_{i=1}^{IK} w_i} \tag{2.3}$$

where w_i is a distance weighting factor and IK is the total number of gages along a selected coastal segment, perhaps a continent. The worldwide estimate of eustatic sea level, $\eta_E(t)$ could then be obtained by averaging over all available coastal segments

$$\tilde{\eta}_E(t) = \frac{1}{IKTOTAL} \sum_{IK=1}^{IKTOTAL} \bar{\eta}_{IK}(t) \tag{2.4}$$

Other ways of extracting meaningful information relating to post-glacial rebound could include averaging first over longitude for certain increments of latitude.

2.4 RESEARCH NEEDS

In general, improvements in our understanding of eustatic sea level change can come about through use of the existing data base or development of new data. Extraction of more meaningful results from the existing data base will require either more powerful analysis procedures or an improved understanding and application of the physics of relative sea level change,

including the noise present in the records. Enhancement of the existing data base through new measurements will most likely occur through satellite altimetry once this is proven to centimeter accuracy over the open ocean. Additionally, in some cases much can be learned locally about anthropogenically generated compaction in areas of tide gages through the installation of rather simple compaction measurement devices. One feature of new data is the length of time that will be required for such data to "mature" to yield significant meaningful information.

2.4.1 Use of Existing Data

Analysis in light of the physics of RSL change appears to be the most effective and productive use of existing data. In particular, accounting for the contribution of long period waves as explored by Sturges (1987) would allow interpretation and removal of a major portion of the noise in the RSL measurements.

A second productive area is a more thorough analysis than presented previously of the contribution of post-glacial adjustment of the earth following the last ice age. As noted previously, Lambeck and Nakiboglu (1984) have inferred from viscous models of the earth that the actual eustatic rise is roughly one-half to two-thirds the value determined from analysis of records based only on areas of relative stability. Improved estimates of eustatic sea level rise could be based on either a more inclusive data set with or without the use of a viscous earth model. Obviously more meaningful results could be obtained with the combined approaches simultaneously. The approach envisioned here is in general the same as applied in "physical principles" with the addition that the global viscous model would be employed for interpretation, guidance and confirmation of the results obtained.

Most approaches of direct analysis attempt to reduce the noise in a record on a station-by-station basis through determining some sort of RSL estimate through fitting to the data. Unfortunately, the noise in individual records is such that at least 20 to 40 years of data must be available at the individual gages before these results can be considered meaningful. An approach that would make these results meaningful early after their availability is the weighted averaging of many stations along a

coastline to establish a more stable value. This averaging length could encompass, for example, the North American or North and South American shoreline(s). Thus, if a wave with length exceeding the expanse of the stations encompassed were contributing to the "noise," this process would tend to reduce or (in the very fortuitous cases) eliminate its contribution. By first averaging over long segments of the shoreline, weighting each station by its alongshore influence length, then combining appropriately the results for various such shoreline segments, a much more stable year by year value could be obtained (i.e., Eqs. 2.3 and 2.4). This would allow effective use of such data as are available for the east coast of South America where eight of the twelve available gages are less than 30 years in duration. As is evident from Fig. 2.6 which presents the mean annual sea level variation of Pensacola, Florida, 30 years is not adequate to obtain a stable estimate from an individual gage.

2.4.2 Need for New Data

There are two types of new data that would contribute to improved estimates of ESLR: those that contribute immediately and those that would require a data base of at least several years before meaningful results could be obtained. It is anticipated that even with the potential benefits of satellite altimetry, at least one decade and possibly two decades will be required before adequate confidence will be placed in these data to yield accepted reliable estimates of eustatic sea level rise. Three research needs in the category of "new data" are described below.

Compaction Gages - As is well-documented by a number of studies, withdrawal of ground water and hydrocarbons can contribute to substantial subsidence and thereby a "relative sea level rise" (see also section 2 for a discussion of compaction effects). It is worth noting that this is probably the only component that realistically can be controlled by humans. The obvious general but not universal correlation of areas of tide gage locations and ground fluid extraction near population concentrations justifies a possible concern over this activity. Also the fact that these are the areas that continued RSL rise may contribute most to the ultimate response cost (relocation, defense, repair, etc.) makes it important that

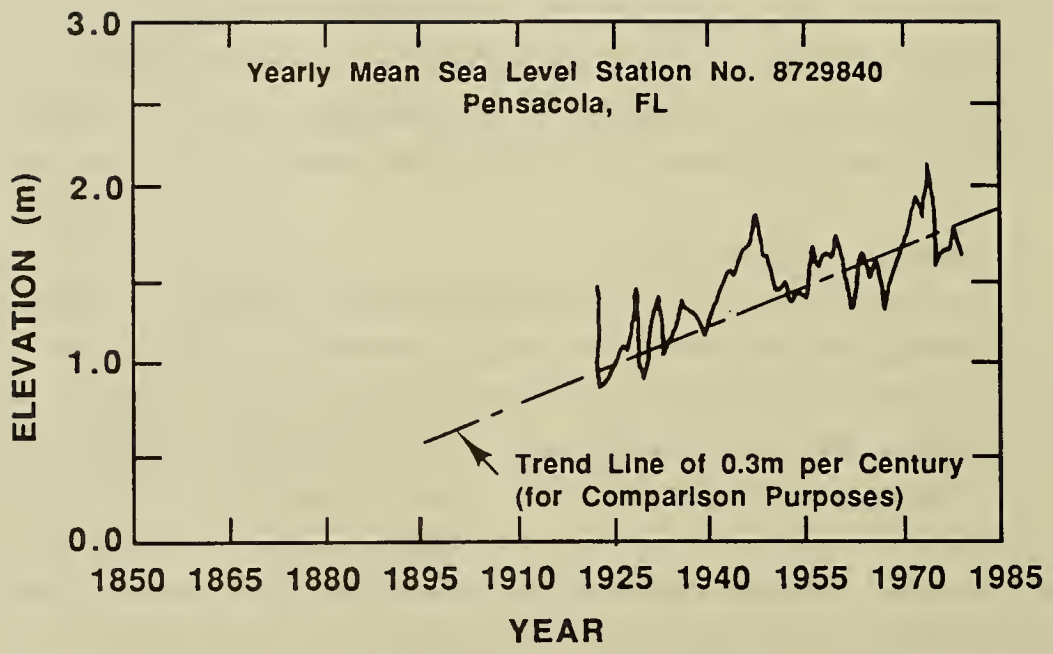


Fig. 2.6. Average Annual Sea Level Variations for Pensacola, Florida (adapted from Hicks et al., 1983).

the significance of anthropogenically induced subsidence be quantified and possibly controlled as early as possible.

Very simple and sensitive compaction meters have been utilized in quantifying this effect in the vicinity of Osaka and Niigata, Japan among other locations. A schematic of two such gages is presented in Fig. 2.7. Each installation consists of an outer casing lining a hole drilled to some depth, h. The inner pipe of slightly smaller diameter is founded on the stratum at depth h. Thus the relative vertical movement between the top of the inner pipe and the general ground level represents the total compaction over the upper sediment column of thickness, h. To establish differential compaction, several such devices would be required at each location of interest. Ideally installations would be made near tide gages and also remote from cities but say inland and in the same geological formations as those near the tide gages. These gages would commence yielding valuable data immediately, and it may be possible to supplement the compaction data collected with models using data representing the geological formations and the history of past ground fluids extraction to estimate earlier compaction. Such results would be invaluable in providing more reliable estimates of past and future eustatic sea level rise.

New Tide Gage Data - Referring to Figs. 2.3 and 2.4a, it is clear that the southern hemisphere is especially deficient in long-term tide gage data. A number of relative short-term tide gage records are available along the east and west coasts of South America; however, there needs to be an effort on an international basis to install and maintain additional gages to provide a representative distribution. In addition to the southern hemisphere, more insular tide gages and tide gages along the open coast are needed. A first phase effort could be a survey to identify such sites.

Satellite Altimetry - This new technology should soon yield absolute vertical accuracies of centimeter accuracy. Thus, sounding much of the ocean surface would allow much broader coverage and very importantly does not require reliance on coastal measurements. It would appear appropriate to continue a dense network of tide gages for sea level rise purposes for

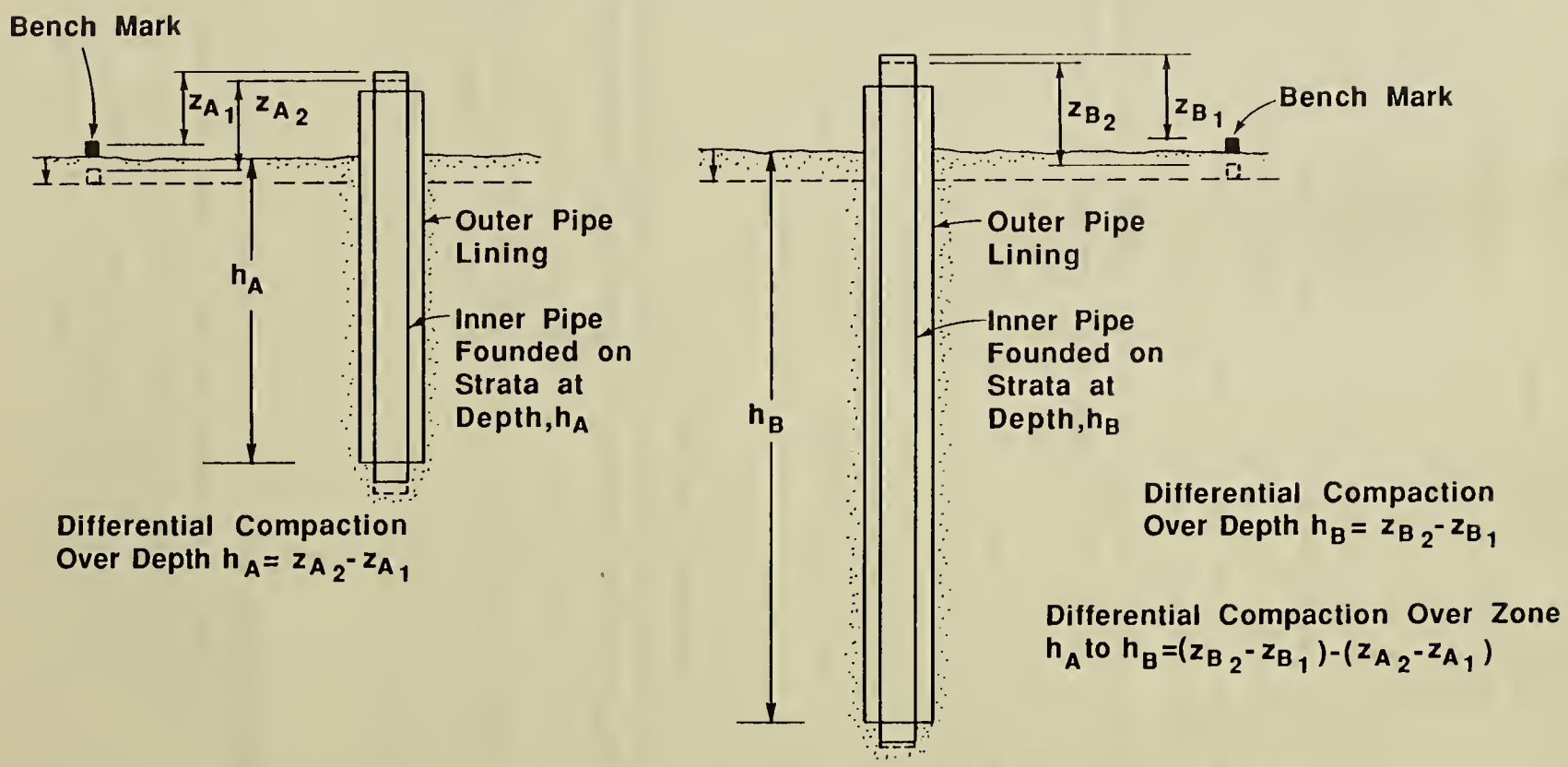


Fig. 2.7. Use of Two Compacting Gages to Obtain Compaction Distribution over Depth Zones h_A, h_B, and h_B-h_A.

several decades after such accuracy is claimed to assure that future needs will be met, and also to allow comparison of the broader satellite coverage and the long-term tide gage results.

3. COMPACTION EFFECTS

3.1 INTRODUCTION

Compaction results in the subsidence of ground level due to reduction in the void ratio of the underlying soil, and in coastal areas contributes to a local relative rise in sea level. Reduction in void ratio is often the natural response of a soil to an increase in loading, because an increase in the interstitial stresses between solids is required. An increase in the loading of a soil stratum can be the result of an increase in loading on the ground surface (e.g., building construction or additional sediment deposition), or due to removal of ground fluid (e.g., water, oil, or natural gas). Compaction occurs in nature as mud is deposited on the beds of rivers and estuaries, and especially in river deltas. Another example is the increase in loading as a barrier island migrates over a stratum of peat, causing the peat to compact and ground level to subside. Because compaction is a time-dependent process, the relative rate between deposition and compaction will determine whether bed elevation increases or decreases. Compaction of a region can also be induced by man, due to 1) loading by the weight of structures, 2) the extraction of oil and natural gas, and 3) depletion of the groundwater table due to active pumping or by preventing recharge of aquifers.

The literature in soil mechanics and foundation design is too replete with articles on the general topic of compaction to review in detail. The proceedings of a symposium "Land Subsidence" held in Tokyo in 1969 (in reference list in section 12) provides a thorough treatment of the causes of compaction, its theoretical description, field measurement techniques and analysis, physical consequences and remedial measures. Much of the subsequent material is gleaned from this collection of studies. The inverse problem (i.e., the effect of sea level rise on compaction and subsidence) has received only limited attention, most recently by Chappell et al. (1982). However, their calculations indicate that a sea level rise on the order of tens of meters would be required for noticeable subsidence of the ocean floor and possible raising of the adjacent land mass.

Shiffman et al. (1985) review the available theories regarding consolidation (compaction). The simplest is Terzaghi's "Conventional Theory" governed by

$$c_v \frac{\partial^2 u}{\partial z^2} = \frac{\partial u}{\partial t} + \frac{\partial u_o}{\partial t} \quad \frac{\partial \sigma}{\partial t} \tag{3.1a}$$

$$cv = \frac{k(1 + e_o)}{\rho_w a_v} \tag{3.1b}$$

where u is the excess pore water pressure, u_o is the hydrostatic pressure, σ is the total stress applied to the system, k is the hydraulic conductivity, e_o is the initial void ratio, ρ_w is the mass density of the fluid (water), and a_v is the compressibility of the soil skeleton. Solving Eq. 3.1 for u and applying the continuity equation for conventional theory

$$\frac{\partial}{\partial z} \left(\frac{k}{\rho_w} \frac{\partial u}{\partial z}\right) = \frac{\partial n}{\partial t} \tag{3.2}$$

soil porosity n is determined. Knowing the porosity as a function of time and the initial thickness of the soil layer, the time history of ground level subsidence can be calculated. Except for very idealized cases, this problem must be solved numerically. Shiffman et al. (1985) also describe a nonlinear finite strain theory, which removes several assumptions of conventional theory but requires difficult numerical solution. Fig. 3.1 displays comparison of the two theories to centrifuge experiments, with the finite strain theory providing good results.

3.2 MEASURING COMPACTION

As noted in section 2, a simple yet effective device for measuring compaction rates has been developed in Japan and has been widely used there for at least the past 30 years, see Murayama (1970). This device, shown in Fig. 3.2 (see also Fig. 2.7), consists of two concentric pipes that penetrate to a desired non-compactable stratum. The outer pipe is perforated to allow the groundwater table to move freely up and down in the

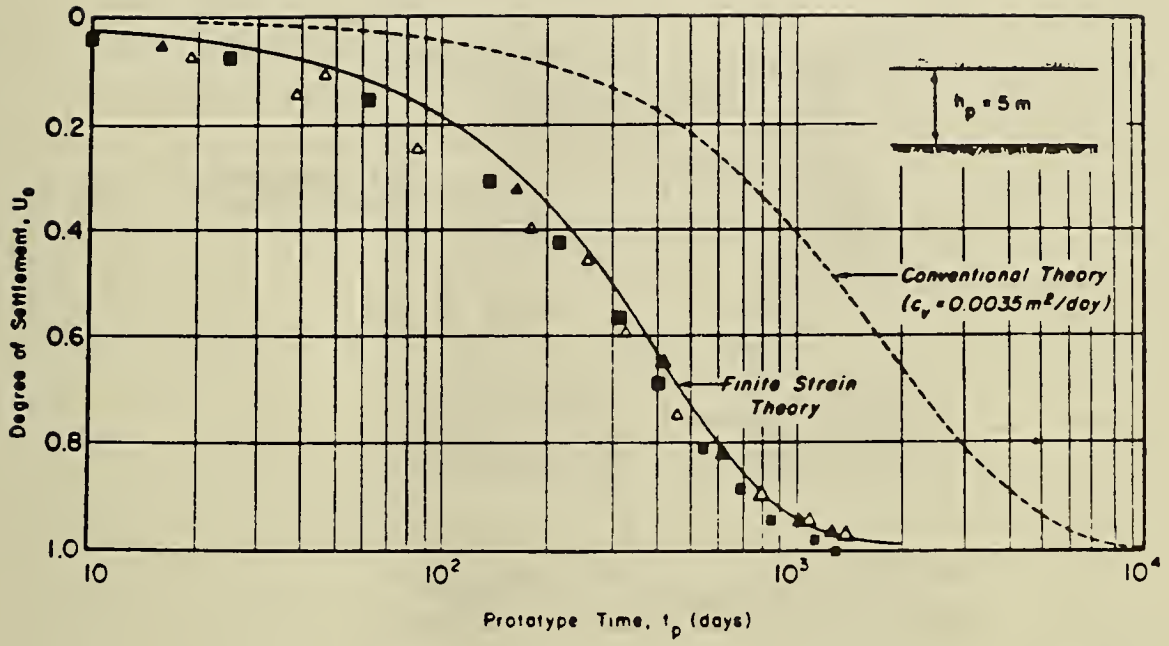


Fig. 3.1. Results of Centrifuge-aided Compaction in Comparison to Two Theories (after Schiffman et al., 1985).

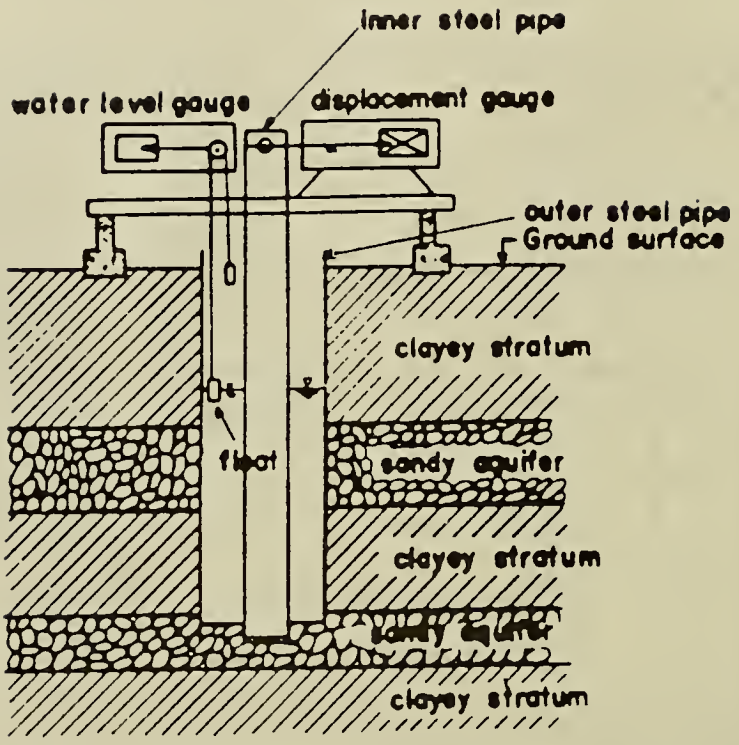


Fig. 3.2. Device for Monitoring Compaction and Groundwater Elevation (after Murayama, 1970).

casing. A float-type gage monitors the water level. A strip chart and pen displacement gage, mounted on a foundation that "rides" the ground surface, records the subsidence as the pipes appear to protrude from the ground. Several of these gages located in the same area, but penetrating to different strata, provide information about the vertical distribution of compaction. A single gage which penetrates to bed-rock will record the total subsidence.

3.3 IMPLICATIONS OF COMPACTION

Compaction enters the discussion of sea level rise in two distinct places. First is the obvious effect that relative sea level will rise as ground or bed level subsides, resulting in deeper water in rivers and estuaries, and increasing the likelihood of erosion and flooding in coastal communities. This will occur even without global sea level changes and seismic activity. Second is the possible contamination of estimates of eustatic rise due to compacting of regions where tide gages are located. Gornitz (personal communication) indicated that 23 tide gage stations located on compactable coastal plain sediments along the southern East coast of the U.S. indicate a sea level rise of 2.96 ± 0.78 mm/yr as compared to 2.57 ± 0.71 mm/yr for 16 stations in the north located on crystalline rock. Although the means do not differ significantly, the large variance in each may well be due to compaction and tectonic effects. Compaction rates comparable to estimates of eustatic sea level rise (~ 1 mm/yr) are not obvious without detailed measurements using devices such as that described. Because tide gages are usually located near coastal cities where both loading by structures and groundwater extraction/depletion are to be expected, the potential for compaction contamination of the measurements exists.

3.4 REMEDIAL MEASURES

Of all types of subsidence, only that which is man-induced can be prevented, arrested, and perhaps partially reversed. Extraction of oil and gas can be accompanied by recharge of the soil stratum with water, as was the case at Terminal Island, California to be discussed. Protection of the

surface recharge areas of aquifers, and water use management to avoid extreme draw-down of the water table also can prevent or reduce compaction.

3.5 EXAMPLES

Mississippi River Delta - A striking example of subsidence due to natural compaction is the delta of the Mississippi River. According to May et al. (1983), the Louisiana coast is retreating at an average rate of 4.2 m/yr, most of which is attributed to erosion and inundation in response to relative sea level rise induced by natural compaction. The levees built along the river have cut off the source of sediment to the mud flats, and their natural rate of compaction is causing some areas to sink at rates of 1 cm/yr or more (see also Table 10.1). Only in a small area of delta formation is the rate of deposition greater than the rate of compaction. This high rate of rise in relative sea level is drowning salt marshes and causing existing small sandy barriers to migrate over the backbarrier muds, further exacerbating the compaction. Penland et al. (1985) predict that at present rates of sea level rise, the Chandelieur Islands and Isles Dernieres will be lost during the next 100 years. Because the loading in this region is naturally induced and the affected area so large, the only functional remedial measure would be to remove the levees in the delta region in hopes of restoring the sediment supply and deposition rate. Although proven successful on a local scale, this is not a cost-effective nor practical solution on a regional basis.

Terminal Island, California - This classic example of the increase in relative sea level due to man-induced subsidence demonstrates many of the possible consequences of natural sea level rise. Due to withdrawal of oil and gas from the Wilmington Oil Field, an area 5 km wide and 6.5 km long subsided an average of about 1.5 m, and encompassed Terminal Island and a portion of Long Beach, California. In some areas the overall subsidence reached 7 m and resulted in considerable damage to harbor facilities as relative sea level rose. This damage required substantial remedial efforts including diking in areas of extreme subsidence, reconstruction of damaged facilities, and bridge repair. The compaction was arrested by injecting water into several of the existing wells in order to maintain pore pressure

as the production wells continued operation. It should be stressed that the rate of increase in relative sea level in this instance was much greater than any expected rates due to eustatic or neotectonic changes.

Japan - Several regions of Japan have experienced large rates of subsidence due to compaction, generally caused by overpumping of groundwater. Ground elevations in Niigata Prefecture and the cities of Osaka and Tokyo have dropped as much as 4 m in the past 40 years, sometimes reaching rates as high as 16 cm/yr (Takeuchi et al., 1970). Fig. 3.3 displays the isolines of the total amount of land subsidence in Osaka from 1935 to 1968. The subsidence is greatest near the coast (280 cm) and small (40 cm) in the hilly region in the center of the city where the compactible stratum is thin. Fig. 3.4a displays monthly measurements of groundwater elevation and Fig. 3.4b shows the corresponding monthly rates of compaction. The two are clearly correlated. The period where subsidence stopped is due to destruction of the city during the bombing of World War II when pumping of groundwater ceased. The installation of an industrial water system and the reduction in pumping started in 1961 have since raised the groundwater table and arrested the subsidence. In Niigata the most severe subsidence has also occurred right on the coast. In all of these cases, regulations controlling groundwater pumping have since been enacted, plus recharge has been practiced in several of the regions where the subsidence is particularly acute. These measures have always proved successful in at least slowing the rate of compaction. In the Tokyo region however, 253 km of embankments, 41 sluice units, and 9 pumping stations were required to protect against typhoon flooding and extreme tides, and to provide drainage for rainwater (Ukena et al., 1970; Tagami et al., 1970). These are precisely the types of measures that may be required in many coastal cities within the next century.

3.6 RESEARCH NEEDS

One important aspect of compaction that requires investigation is its effect on the tide gage measurements used to determine sea level rise, as noted in section 2. Although gage elevations are often surveyed in relation to bench marks that are anchored to bedrock, the error inherent in

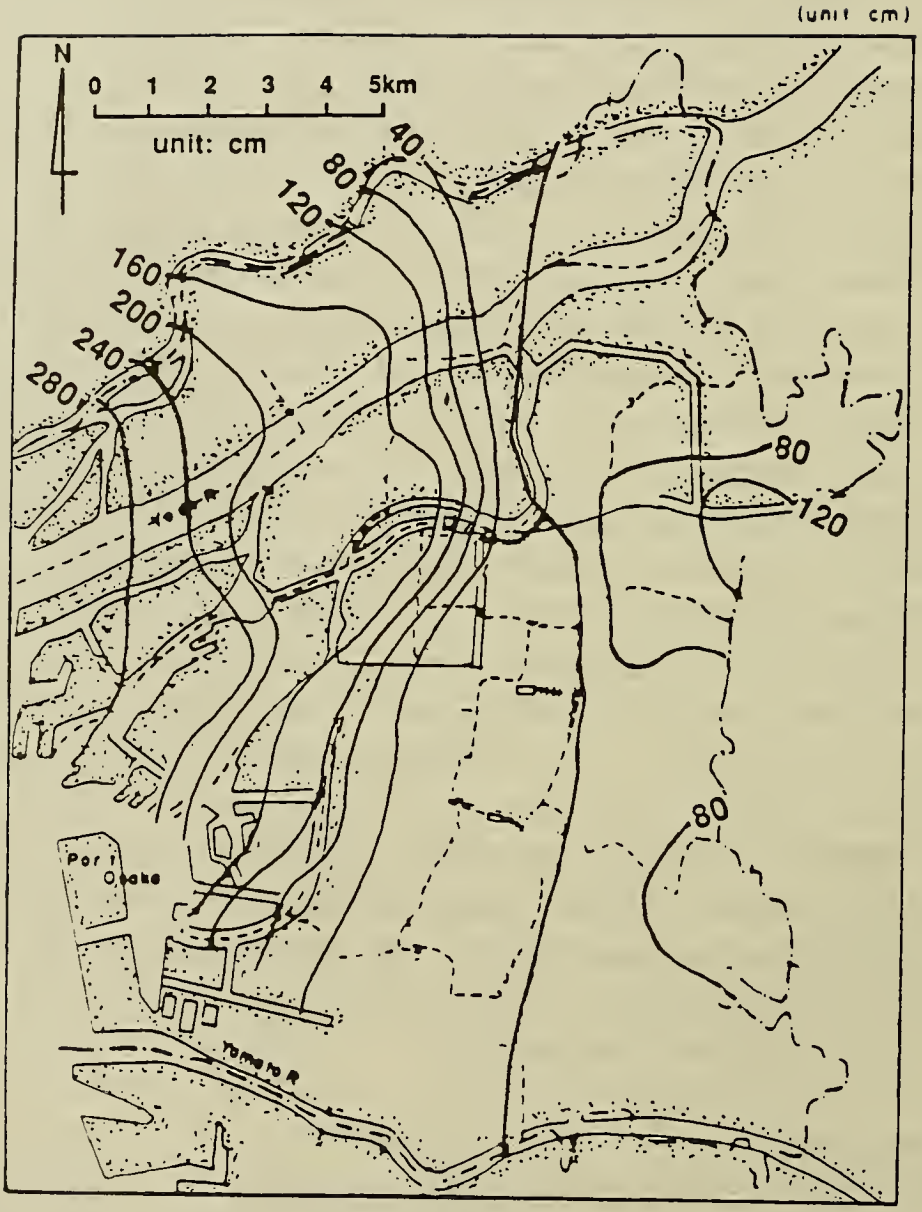


Fig. 3.3. Isolines of Total Subsidence (in cm) from 1935-1968 in Osaka, Japan (after Murayama, 1970).

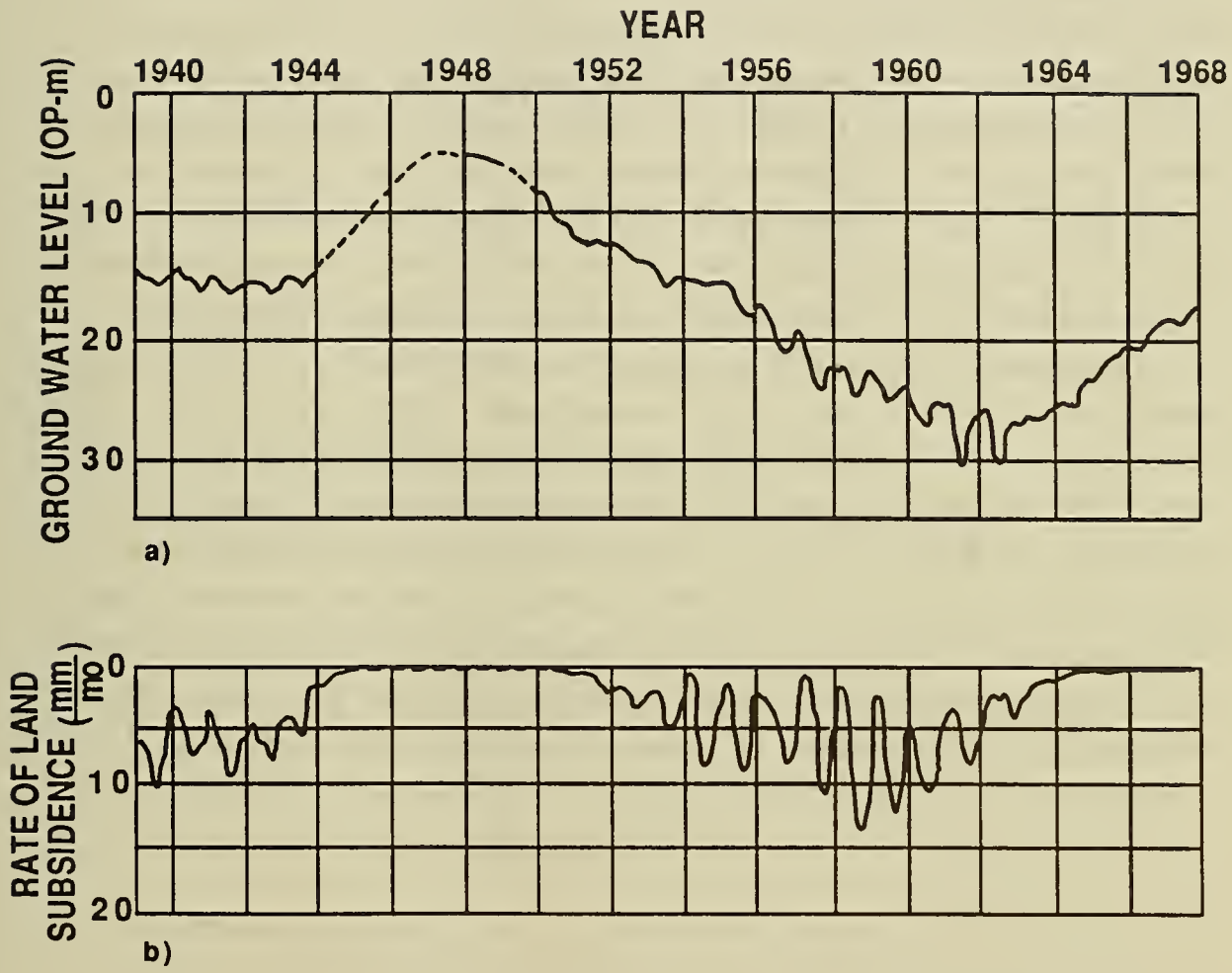


Fig. 3.4. Monthly Record of a) Groundwater Level and b) Rate of Subsidence in Osaka, Japan (after Murayama, 1970).

leveling over possibly long distances would favor a more direct indication of any local compaction. It is recommended that a few experimental groundwater table/compaction devices be installed near selected tide gages. These would be located in communities where demand for the local groundwater is high, and compressibility of the underlying strata significant. If these devices prove useful, more should be added until, ideally, every tide gage used in making sea level rise estimates has at least one accompanying compaction device.

In many regions, tectonic activity may be equal to or greater than compaction. The only means to resolve the apparent rise in sea level at a tide gage into its subsidence, tectonic, and eustatic components with a degree of confidence is to relate tide and compaction gage elevations to an ultra-precise geodetic reference system. Such a program utilizing Very Long Baseline Interferometry (VLBI) and the Global Positioning System (GPS) is described by Carter et al. (1986). The GPS or other satellite system should also be used to monitor the deep ocean as stated in section 2.

A program is also needed to document compaction rates in those coastal areas currently experiencing high rates of erosion and shoreline retreat to see if compaction is playing a role, and to determine if remedial measures can be implemented. Installing arrays of compaction measuring devices will also permit study of the long-term behavior of the local subsidence as well as its relation to fluctuations and secular depletion of the water table. By sinking nearby devices to different depths, the vertical distribution of compaction can be determined, and by placing arrays farther inland along a transect the spatial behavior of compaction rates can be studied.

4. TIDAL RANGE EFFECTS

4.1 INTRODUCTION

The effect of sea level rise on the open coast and estuarine tidal ranges is a matter of significance as far as the dynamics of shoreline response is concerned, including such processes as coastal flooding, salinity intrusion and sediment transport. An obvious question is whether a rise in the range, should it occur, would overshadow the effect of the mean sea level rise itself. The phenomenon is strongly site-specific, depending upon local morphological and meteorological conditions, and also on remote forcing due to macro-scale oceanographic phenomena.

Astronomical tides are shallow water waves even in the deepest ocean, and therefore "feel the bottom." Conversely, therefore, the bottom topography and frictional resistance influence tide propagation in the sea. Since shorelines define the boundaries of the offshore shelf which is usually quite "shallow," nearshore tides are strongly influenced by the shelf topography. The distinction between tide measured "along the open coast" and, for example, at a bay entrance therefore becomes somewhat blurred. With reference to tide measured inside a bay as opposed to outside, Mehta and Philip (1986) noted that "the definition of 'outside' remains somewhat obscure in physiographic terms...." However, they added that "restrictive dimensions of bays compared to the sea impose water level oscillations whose range and frequency may be partially unrelated to oscillations outside." Furthermore, from the point of view of organizing data, the distinction between open coast tide and bay or estuarine tide may be retained, as in the following description.

4.2 LITERATURE REVIEW

The principal tide-generating forces arise from the gravitational pull exerted on the earth's surficial water mass (and to a much smaller extent on the entire mass of earth; see, for example, Hendershott, 1972) by the moon and the sun. Darwin (1898) presented an "Equilibrium Theory of the Tides," which rovides a useful qualitative description of some of the main features of the tide phenomenon based on a force balance involving

gravitational attraction and centrifugal reaction for the system comprising the earth, the moon and the sun. This theory has been summarized by Dean (1966); it highlights the role of the basic forces in generating periodic oscillations of the water surface, and their dependence on such factors as the latitude, the declination of moon and the relative effects of the moon and the sun.

During the 1920s, Proudman (see, for example, Proudman, 1925) published a series of articles in which he investigated various aspects of tidal motion including the Coriolis effect due to earth's rotation. The significant advance made relative to the equilibrium theory was accounting for the actual motion of water particles on the rotating earth. Computer technology has now made it feasible to simulate tidal motion over entire oceanic masses. Early computations were based on solutions of Laplace's tidal equations (LTE). A review of numerical models of the sixties and the seventies has been provided by Hendershott (1977). Subsequently, more general forms of the Navier-Stokes equations of motion have been solved. A recent review of solutions of these ocean tidal equations (OTE) has been provided by Schwiderski (1986).

Tides in the nearshore environment are considerably influenced by winds, waves, bottom topography as well as temperature- and salinity-induced stratification. Where astronomical tides are small, e.g., along U.S. Gulf coast, non-tidal forcing often assumes overwhelming significance and modeling of a purely deterministic nature becomes difficult. Physical considerations along these lines have been reviewed by Csanady (1984).

Proudman's contributions also included considerations for tidal motions in channels of various cross-sectional shapes, and the effect of coastal configuration on offshore tidal features. A good review of simple analytic approaches for tidal propagation in estuaries, without and with bottom frictional effects, has been presented by Ippen and Harleman (1966). For the fundamentals on numerical methods for estuarine hydrodynamics, the works of Dronkers (1964) and Abbott (1979) may be cited. Nihoul and Jamart (1987) have edited a series of contributions on the state-of-the-art modeling techniques of marine and estuarine hydrodynamics using three-dimensional numerical approaches.

A special class of tidal hydraulics pertains to the hydraulics of tidal inlets or entrances connecting the sea to relatively small and deep bays. A simple, coherent theory for predicting water level variation in the bay for a given, sinusoidally forced, sea tide has been presented by Keulegan (1967). Mehta and Özsoy (1978) have reviewed various approaches including developments previous and subsequent to Keulegan's contribution.

4.3 PHYSICAL PRINCIPLES

4.3.1 Tidal Propagation

According to the equilibrium theory of tides, the tidal amplitude can be shown to be proportional (to leading order) to the fourth power of earth's radius, considering the moon-earth system. Since this number (6,378 km) is so large compared to any expected effect of sea level rise (i.e., increase in earth's radius), the corresponding change in the tidal range on this account would be negligible. In order to evaluate the effect of sea level rise on the tidal range, the nature of propagation of tide in very shallow waters must be considered.

The simplest description of tide in the dynamic sense is that of a shallow water wave moving along the x-direction with a speed or celerity, C_o. If a frictionless bottom is assumed, the wave equation is

$$\frac{\partial^2 \eta}{\partial t^2} = C_o^2 \frac{\partial^2 \eta}{\partial x^2} \tag{4.1}$$

where $\eta(x,t)$ is the instantaneous water surface elevation. The celerity, $C_o = (gh)^{1/2}$ where g is acceleration due to gravity and h is water depth.

The effect of friction can be accounted for by including an additional term on the right hand side of Eq. 4.1. Thus, for example, this term under the assumption of linearized friction is $-gM\partial\eta/\partial t$, where M is an empirical coefficient accounting for the magnitude of bottom friction. Friction slows down the speed of propagation (celerity), decreases the current speed and reduces the tidal range compared with frictionless tide. The effect is depth-dependent, and it can be shown that in fact it varies with $h^{-1/3}$ which means that increasing the water depth would decrease frictional damping, thereby increasing the tidal range. Observations in the German

Bight (southern North Sea) suggest this type of a trend, as will be noted later.

Within the estuary itself, increasing the water depth can have a drastic effect on the tidal range. The majority of present day estuaries are of holocene origin, having been formed since the last ice age and accompanying sea level rise. In some, sea level rise has caused the depths to increase while in others, sedimentation rates have been high enough for the depths to have "kept pace" with sea level rise. In a few cases, for example, some estuaries in China (e.g., Qitang), sedimentation rates have essentially exerted an overwhelming control, causing the depths to decrease in spite of sea level rise, and thereby pushing the mouth seaward.

While increased water depth would generally increase the estuarine tidal range, the opposite effect could occur, for example, in cases where tidal resonance is a significant factor. This can be illustrated in a simple way by considering the case of a tidal wave entering a frictionless channel closed at the upstream end. In this case, considering complete wave reflection at the closed end, the incident and reflected progressive waves combine to form a standing wave, as shown in Fig. 4.1. The estuary is of length ℓ, with the closed end at $x = 0$ and the mouth at $x = -\ell$. If the range of the progressive wave is H, the range of the standing wave at the closed end will be 2H. The standing wave envelope is thus defined by an antinode at the closed end and a node in the sea. It can be shown (Ippen and Harleman, 1966) that the ratio, R, of the amplitude, η_{om}, at the closed end to the amplitude, $\eta_{-\ell m}$, at the mouth will be (ignoring bottom friction)

$$R = \frac{\eta_{om}}{\eta_{-\ell m}} = \frac{1}{\left|\cos\left(\frac{2\pi\ell}{L}\right)\right|} \tag{4.2}$$

Since $\left|\cos(2\pi\ell/L)\right| \leq 1$, in general, the tide at the closed end will be higher than that at the mouth. This type of a resonance effect is well known, and occurs in such estuaries at the Bay of Fundy, Canada, and at Cambay in India. Given such a behavior, a situation can arise whereby an

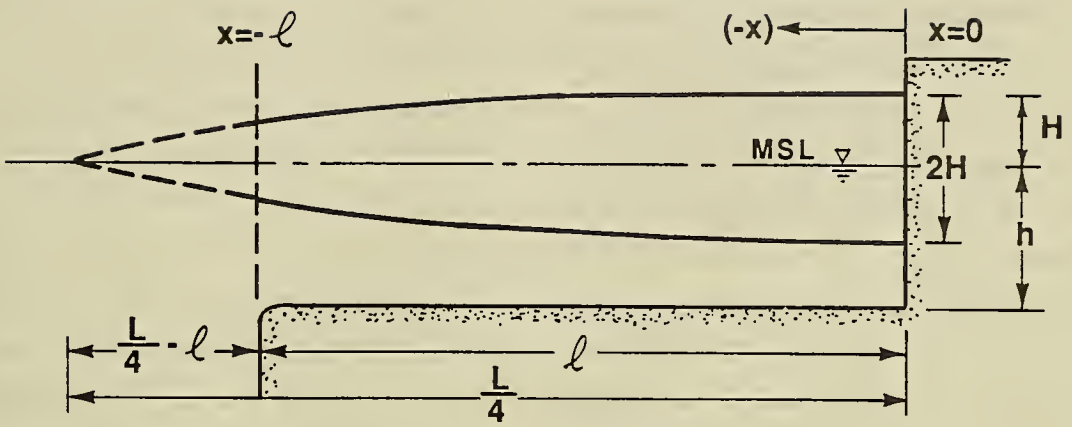


Fig. 4.1. Tidal Wave Envelope in an Estuary in which the Wave is Reflected at the Upstream Closed End.

increase in water depth would in fact decrease the difference between the tide at the closed end and that at the mouth.

Consider first the case of an estuary of mean water depth, h = 15 m. Given an estuary length, ℓ = 108 km, from Eq. 4.2 R = 3.7, for a semi-diurnal tide. Now if h is increased, for example, by 2 m, R is reduced to 2.60 (assuming no change in the estuary length). Further suppose that as a result of the 2 m sea level rise, the tidal range at the mouth increases by 10%, say from 1 m to 1.10 m. Then, by virtue of Eq. 4.2, the range at the closed end will decrease, from 3.2 m to 2.9 m.

A bay-like water body connected to the sea via an entrance will experience range amplification as the frequency of tidal forcing approaches the natural period of oscillation of the water body. The situation is analogous to the response of a damped harmonic oscillation (Mehta and Özsoy, 1978). In a number of bays along the U.S. coastlines, for example, the tidal range in the bay is greater than that outside (O'Brien and Clark, 1974). Amplification becomes most pronounced when the forced and natural frequencies are equal. If therefore an increase in water depth due to sea level rise were such as to shift bay response away from resonance, the tidal range relative to that at the mouth could, as illustrated previously, decrease in spite of the opposing trend caused by decreasing bottom friction and increasing tidal admittance with increasing water depth. In a great many inlet/bay systems, however, bottom friction in the inlet channel controls the bay tide; hence in these cases resonance is not a critical factor.

4.3.2 Superelevation Effect

In most bays, the tidal mean water level is usually different, often higher than mean sea level. The difference, referred to as bay superelevation, results from a number of physical factors. Mehta and Philip (1986) reviewed these factors, and the physical mechanisms by which they generate superelevation. Representative maximum superelevations corresponding to each cause, as might be found from measurements, were suggested; Table 4.1 gives a summary of the findings. Among the listed causes, sea level rise will directly or indirectly influence inlet/bay geometry, sea tide, salinity, wave penetration and some other factors.

Table 4.1. Representative Bay Superelevations (after Mehta and Philip, 1986)

Cause	Superelevation[a] (cm)	Comment
Inlet/Bay Geometry	5-30	Effect of shallow bar is more important than changing geometry with tide; hence seasonal and episodic response
Sea Tide	10	Theoretical estimate; no verification; believed to be a small contribution compared to others
Runoff	50	Major factor; strong seasonal variation
Salinity	15	Important in estuaries rather than bays (no runoff); seasonal variation
Wind	10-15	Local forcing and remote forcing can both be equally important; seasonal and episodic response
Waves	5-10	Induced pileup behind reefs may be important; seasonal and episodic response
Other Factors	1-30	Modification of tide during upland propagation and Coriolis effect are significant

[a]Only positive values are indicated. Superelevation can also be negative, i.e. lower bay level than sea, e.g. due to offshore wind.

Since these in turn influence the mean bay level, in the evaluation, for instance, of the change in tidal range due to sea level change, the associated change in superelevation must be additionally considered in calculating the net water change.

Mann (1987) examined the superelevation effect resulting from inlet/bay response to tidal forcing. Tide-averaged hydrodynamic equations were developed and it was shown that bottom friction in the inlet channel is the primary cause of superelevation. Stokes drift, tidal current asymmetry and river runoff were identified (in the absence of such effects as those arising from salinity, wind waves, etc.) as the major governing physical processes. Mann considered the case of a small, deep bay connected to the sea via a long inlet channel. The combined effects of tide and superelevation resulting from sea level rise were evaluated, as will be noted in the next section.

4.4 EXAMPLES

Führböer and Jensen (1985) evaluated long-term sea level trends at ten gages in the German Bight. The evaluation was based on records obtained over a 100-year period from 1884 to 1983. Trends relevant to the present purpose may be illustrated by considering four gages, at Norderney, List, Cuxhaven and Bremerhaven, shown in Fig. 4.2. Of these, the gages at Norderney and List may be considered as "open coast" gages, while Bremerhaven is decidedly up estuary (Weser). Cuxhaven is at the mouth of estuary (Elbe). Table 4.2 gives relevant results.

The rate of change of mean tidal range has been calculated in two different ways for each location. The first is the average rate based on the entire 100-year period (no values were computed for Norderney and List due to insufficient data). The second is based on the last 25-year (= N) record, converted to an equivalent 100-year rate. Comparing Norderney and List to Bremerhaven, it is observed that the tidal rise (N = 25) has been far more significant (threefold) within the estuary than on the open coast. The rise at the estuary mouth is intermediate in magnitude. One likely reason is the effect of reduced bottom friction due to sea level rise. This effect is more pronounced in the shallow estuary than in the deeper sea.

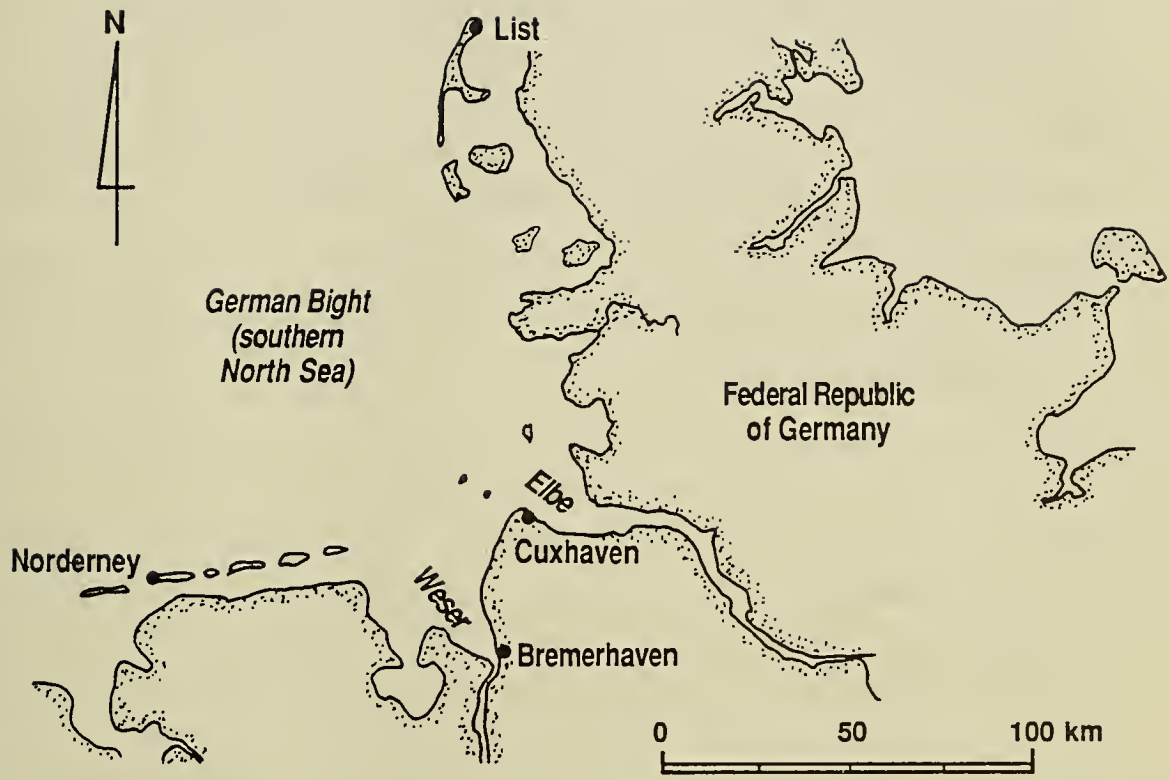


Fig. 4.2. Locations of Four Tide Gages in the German Bight.

Table 4.2. Secular Trends in Mean Tidal Range in the German Bight (modified from after Führböter and Jensen, 1985)

Location	Rate of Change of Mean Range (m/100 yr)	
	N = 100	N = 25
Norderney	-[a]	0.431
List	-[a]	0.369
Bremerhaven	0.380	1.293
Cuxhaven	0.065	0.949

[a]Insufficient data.

It is also interesting to observe from Table 4.2 (for Bremerhaven and Cuxhaven) that the increase in tidal range has been considerably more significant in recent years (N = 25) than what is obtained based on a 100-year record (N = 100). At Bremerhaven, the mean tidal range 100 years ago was ~3.30 m. Thus the range increased there by ~9% during the subsequent 75 years. During the next 25 years the range increased again by about the same percentage.

Führböter and Jensen noted a trend of rising tidal range approximately over the past century at all ten locations examined. They concluded that this trend is not due to any long-term changes in meteorological conditions, but is possibly due to the morphology of the North Sea, a very shallow water body in which the global rise of the mean water level effect is amplified via a standing wave effect. This possibly suggests a situation in which the natural frequency of the water body approaches the tidal forcing frequency with increasing water depth and changing boundaries.

Mann (1987) theoretically simulated the response of inlet/bay systems of assumed geometries to a total sea level rise of 1.3 m, corresponding to a 0.3 m rise over the past century and a 1.0 m projected rise. The bay was assumed to be relatively small and deep, with a surface area of 5×10^6 m^2. The inlet channel was 1,800 m long and 150 m wide. It is illustrative to consider here the case of an initially 1.5 m deep channel. For this shallow system, the ratio of the (semi-diurnal) tidal frequency to the natural frequency is 0.16, which is $<<1$, thus signifying a friction-dominated (as opposed to resonance-dominated) system.

In Fig. 4.3, the resulting changes in the mean bay level and bay tidal range are shown. A 1.3 m rise in sea level decreased bay superelevation (head above mean sea level), from 0.27 m to 0.11 m. On the other hand, reduced friction resulted in an increased tidal range. Initially, the high water (HW) and low water (LW) amplitudes of tide relative to mean bay water level were 0.28 m and 0.25 m, respectively. The tidal range was thus 0.53 m. After a 1.3 m sea level rise, the amplitudes became 0.66 m and 0.56 m (i.e., range 1.22 m).

These data on the effect of sea level rise enable the determination of the high water level within the bay initially, and following sea level

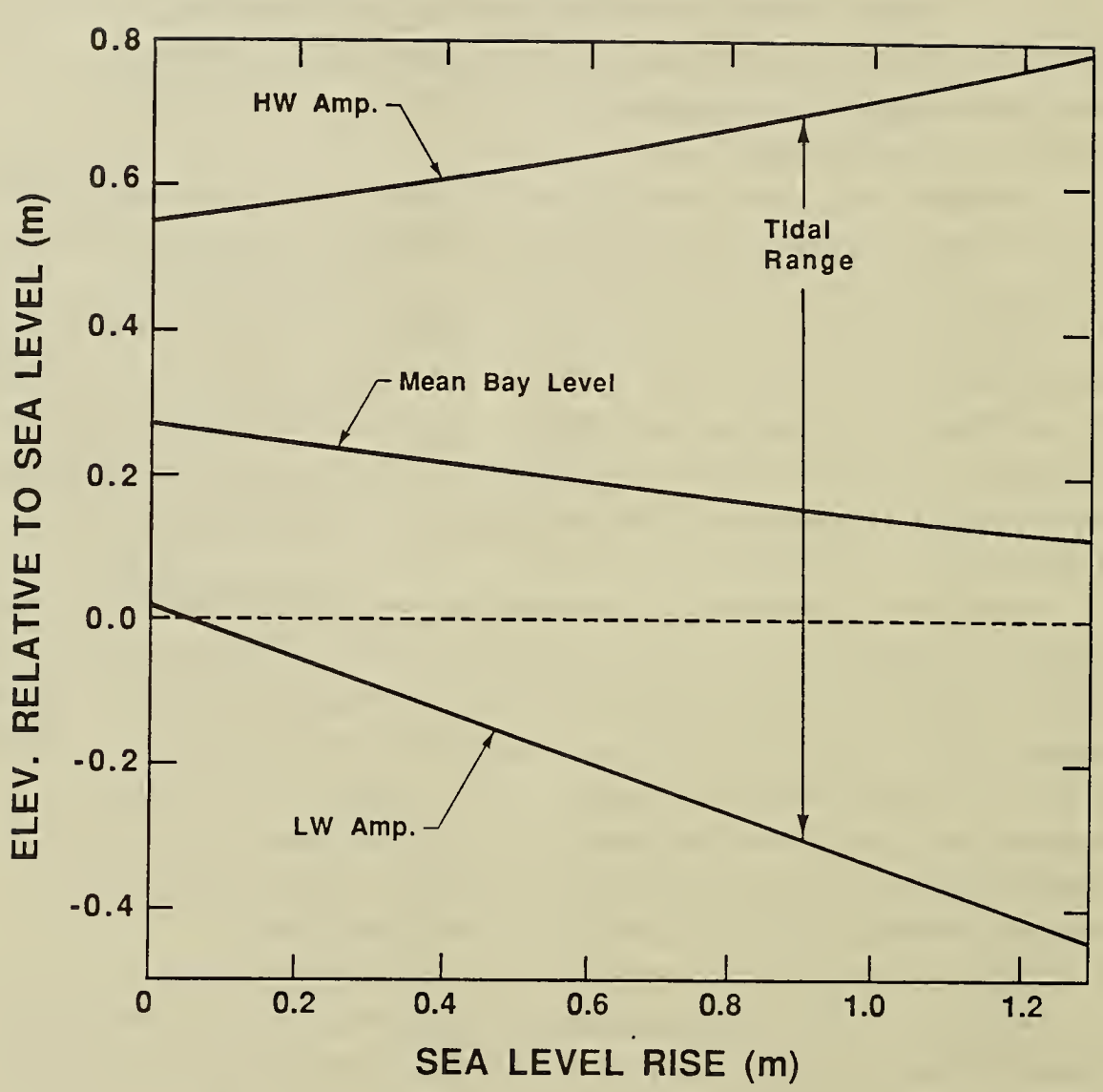


Fig. 4.3. Response of a Shallow Inlet/Deep Bay System to Sea Level Rise: Changes in Mean Bay Level and Tidal Amplitudes (based on computations by Mann, 1987).

rise. Let S = sea level rise, a_{HW} = HW tidal amplitude in the bay relative to mean bay level and B = bay superelevation. Let Δa_{HW} and ΔB represent changes in a_{HW} and B, respectively. Then, initially, the HW level with respect to the initial mean sea level will be $a_{HW} + B$. After sea level rise, it will be $S + a_{HW} + \Delta a_{HW} + B + \Delta B$. Note that in the example considered, ΔB is a negative quantity. Relevant quantities in the present case are: S = 1.3 m, a_{HW} = 0.28 m, Δa_{HW} = 0.38 m, ΔB = 0.27 m and ΔB = -0.16 m. Thus the initial HW level relative to initial sea level was 0.55 m, which rose to 2.07 m subsequent to sea level rise.

The significance of the above result is self-evident; sea level rise could, in addition, increase the tidal range so that, in spite of a decrease in bay superelevation, high water level rise within the bay would become greater than that corresponding to sea level rise alone. It must be pointed out that certain mitigating factors including bottom friction in the bay, ignored in the above example, could lead to a less drastic effect of sea level rise on the bay tide range. Therefore this illustration should be considered to represent an extreme case in terms of quantitative effects.

A noteworthy conclusion based on the result of Fig. 4.3 is that the secular rate of water level rise would be lower in the bay than in the sea, on account of the decrease of bay superelevation. Hicks (1984) selected 19 pairs of gages, one inside the bay and the other at the closest location outside the entrance, for which long-term data were available. For each pair, the difference (outside minus inside) in the secular rate of change of mean water level (mm/yr) was calculated. In 12 cases, this difference was positive, which means a greater water level rise outside than inside the bay. With the exceptions of the Long Branch (NJ)/New York (NY) and Springmaid Pier/Charleston (SC) pairs, where the differences were large (13.1 and 13.6 mm/yr, respectively), the mean of the remaining 10 pairs was 2.6 mm/yr. If bay superelevation changes were the sole effect involved (which is not by any means certain, since the gage data were probably contaminated by any number of physical phenomena), this 2.6 mm/yr change would be indicative of the rate of decrease of superelevation.

Mann (1987) showed that the changes in bay response are greater in shallow inlets than in deep ones. He also found that considering, for

example, the bay to have a gentle boundary slope as opposed to a vertical wall-like perimeter would reduce the changes in superelevation and tidal range compared with the vertical wall case (Fig. 4.3). In general, however, it was concluded that due to an increase in sea level, "additional coastal flooding may occur beyond that due merely to the changes in sea level." Observations by Führböter (1986) in the German Bight estuaries seem to corroborate such a trend.

4.5 RESEARCH NEEDS

Fast computers with large memory storage have made numerical modeling of tides rather sophisticated. In many cases, it seems, modeling capabilities have "outstripped" data quality such that inaccuracies in collected data limit the accuracy of mathematical prediction. Data limitations arise from many causes; it suffices to note two factors.

One pertains to a lack of physical understanding, on a microscale, of phenomena which ultimately affect water level prediction. An example is our understanding of bed forms, the manner in which they change with flow, and the precise relationship between their occurrence and the flow resistance they generate. Such forms may be as small as ripples to large, migratory sand waves found in estuaries and in nearshore waters.

The second factor is related to historic tide records. Many records are highly contaminated by such unaccounted for effects as arising from land subsidence, poor leveling between gages, shifting gage locations, and a general lack of knowledge of the physical surroundings and variations in parameters characterizing these surroundings over the duration of tidal record. Thus, an accurate, quantitative evaluation of superelevation effects would require the deployment of better monitored gages. In addition, Mehta and Philip (1986) noted that our understanding of bay response and its relation to response outside would be considerably enhanced by: 1) establishment of additional primary stations along the open coast, 2) collection of long-term records at several presently designated secondary stations in bays, 3) accurate geodetic leveling connecting additional outside and inside stations, and 4) publication of relevant data in a user-oriented format. National Ocean Service initiated marine boundary programs and tidal datum survey programs appear to be directed

towards this type of effort, particularly with respect to the first three items.

5. STORM SURGE AND WIND-WAVE RESPONSE

5.1 INTRODUCTION

Storm surge is the response of mean water level to the high winds, pressure differential, and rainfall associated with tropical (hurricane) and extratropical (northeaster) storms. The forces which appear to elicit the greatest responses are wind-induced shear, which tends to push water onto the beach, and the inverse barometer effect, which elevates the water level under the eye of a hurricane. For example, Fig. 5.1 displays the observed tides and storm surge associated with Hurricane Carla in the Galveston, Texas area. A complete discussion of all the relevant forces and the equations governing flows induced by storms can be found in the Shore Protection Manual, U.S. Army Corps of Engineers (1984). Solutions to idealized cases are given by Bretschneider (1966a) and Dean and Dalrymple (1984). The dependence of these solutions on nominal water depth will be examined in order to postulate some of the possible effects of long-term sea level rise.

The stress applied to the water by the high winds associated with storms is also responsible for wave generation. Bretschneider (1959) developed a family of curves from non-dimensional significant wave height induced by a hurricane, shown in Fig. 5.2. Wind waves are affected by sea level (water depth) both in their generation and as they propagate over the continental shelf. Shallow water limits the height a growing wave can attain due to steepness-induced breaking and bottom friction, while bottom friction continues to drain energy from the waves as they propagate out of the generation region. Wave generation in shallow water and losses due to bottom friction will be briefly examined in order to identify effects of depth, and hence the consequences of long-term sea level rise. Reference can be made again to the Shore Protection Manual and Bretschneider (1966b) for information on these topics.

5.2 STORM SURGE

For the idealized situation shown in Fig. 5.3a where the continental shelf is uniform in depth, we consider a spatially and temporally uniform

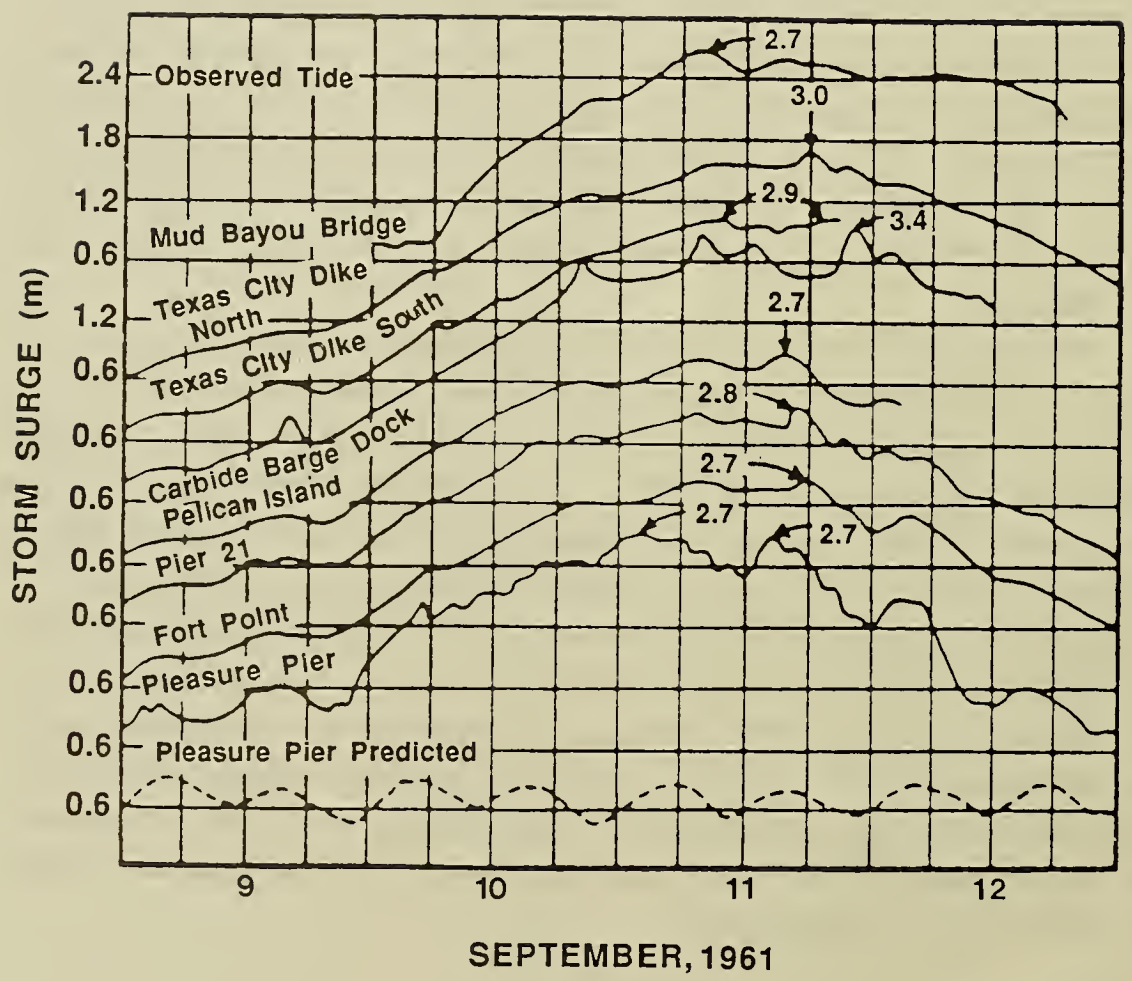


Fig. 5.1. Measured Storm Surge in Galveston, Texas Area during Hurricane Carla (adapted from Army Corps of Engineers, 1984).

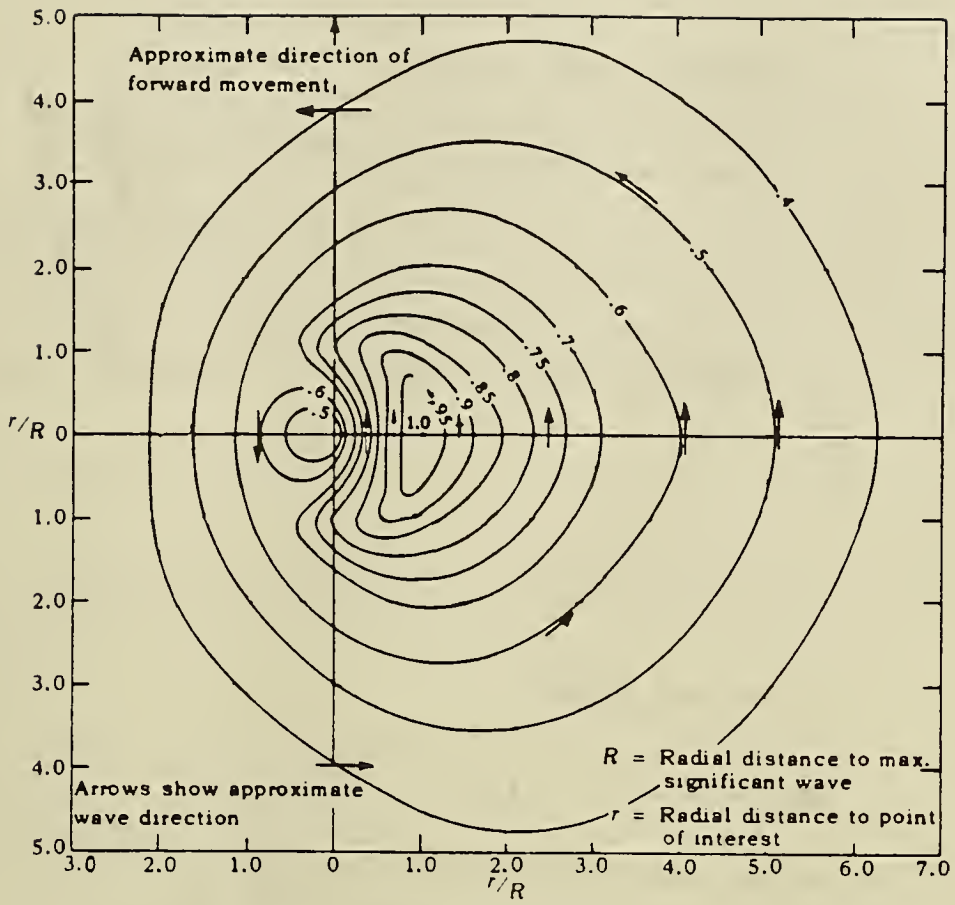


Fig. 5.2. Isolines of Non-Dimensional Significant Wave Height for Hurricane-generated Wind-waves (after Bretschneider, 1959).

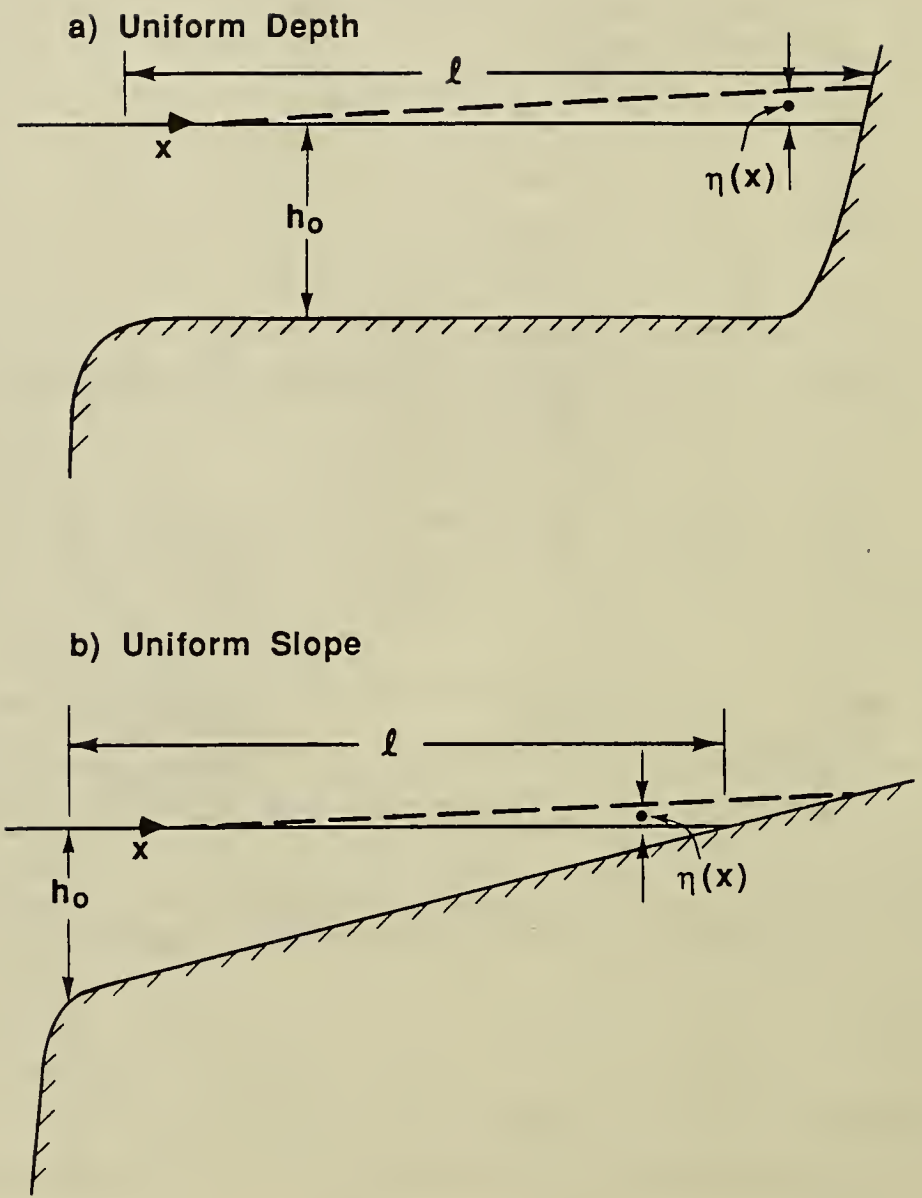


Fig. 5.3. Idealized Geometries for the Continental Shelf: a) Uniform Depth, b) Uniform Slope.

surface shear stress due to the wind associated with the storm. According to Dean and Dalrymple (1984) the set-up, η, for steady-state conditions is given by

$$\frac{\eta(x)}{h_o} = \sqrt{1 + \frac{2B\ell}{h_o^2}\frac{x}{\ell}} - 1 \tag{5.1}$$

where B includes the wind induced shear stress and h_o is the original water depth. Note that at the shoreline ($x = \ell$), the set-up increases with the expanse of the shelf. Rearranging slightly and evaluating this expression at the shoreline yields

$$\eta(\ell) = (h_o^2 + 2B\ell)^{1/2} - h_o \tag{5.2}$$

Taking the derivative with respect to h_o, the dependence of $\eta(\ell)$ on h_o can be examined:

$$\frac{\partial\eta(\ell)}{\partial h_o} = [h_o(h_o^2 + 2B\ell)^{-1/2} - 1] \tag{5.3}$$

$$\frac{\partial\eta(\ell)}{\partial h_o} = \frac{1}{(1 + \frac{2B\ell}{h_o^2})^{1/2}} - 1 \tag{5.4}$$

Because Eq. 5.4 is always negative, η decreases as h_o increases. This indicates that according to this simple model, as long-term sea level rises (h_o increases) the set-up induced by a given wind shear will decrease.

Consider the following situation:

average depth h_o = 10 m

shelf width ℓ = 150 km

average wind

shear "head" $B = 3.3 \times 10^{-5}$ m (wind speed $\simeq$ 12.5 m/s)

and using Eq. 5.2, the wind-induced set-up at the shoreline is calculated: $\eta(\ell) = 0.49$ m. For the same shear and shelf width, but including a 1 m rise in sea level yields $\eta(\ell) = 0.45$ m. There is 4 cm less set-up with sea level rise, but of course the total water level would be 96 cm higher than without the 1 m rise in sea level. It is noted that the conditions in this example correspond to a shallow shelf, such as the Gulf of Mexico. The storm surge on deeper shelves would respond even less to an increase in sea level.

Another relevant idealized geometry for the continental shelf and nearshore region is that of a uniformly sloping bottom, as shown in Fig. 5.3b. Dean and Dalrymple (1984) present an implicit solution for wind-induced set-up

$$\frac{x}{\ell} = \left(1 - \frac{h + \eta}{h_o}\right) - \frac{B\ell}{h_o^2} \ln \left[\frac{\frac{h + \eta}{h_o} - \frac{B\ell}{h_o^2}}{1 - \frac{B\ell}{h_o^2}} \right] \tag{5.5}$$

where

$$h = h_o \left(1 - \frac{x}{\ell}\right) = h_o\left(1 - \frac{xm}{h_o}\right) \tag{5.6}$$

where m is the shelf slope. If we attempt to follow the same procedure as before, $\eta(\ell)$ is given by the transcendental expression

$$0 = \eta(\ell) + \frac{B}{m} \ln \left(\frac{\eta(\ell) - \frac{B}{m}}{h_o - \frac{B}{m}}\right) \tag{5.7}$$

Taking the derivative of Eq. 5.7 with respect to h_o yields

$$\frac{\partial \eta(\ell)}{\partial h_o} = \frac{1 - \frac{B}{m\eta(\ell)}}{\frac{h_o m}{B} - 1} \tag{5.8}$$

from which it is difficult to immediately draw firm conclusions. However, for positive $\eta(\ell)$, from Eq. 5.7 we know that

$$0 < \frac{1 - \frac{B}{m\eta(\ell)}}{\frac{B}{m\eta(\ell)}\left(\frac{h_o}{B}_m - 1\right)} \tag{5.9}$$

Quite sophisticated 2-D (planform) and multiple layer numerical models have been developed since the mid-1950s, which treat more realistic bottom topography and boundary geometry, see e.g., Jelesnianski (1965), Reid and Bodine (1968), Heaps and Jones (1975), Wang and Connor (1975), Wanstrath et al. (1976), Forristall et al. (1977), Chen et al. (1978) and Thacker (1979). The caliber of these models has outrun the quantity and quality of available field data with which to verify them. Complicated numerical models also do not lend themselves easily to examining the general interaction of the forcing mechanisms and surge response, and it may be necessary to look for ways to parameterize and scale the models to extract the effects of sea level rise on storm surge.

5.3 WAVE CHARACTERISTICS

The characteristics of waves generated in deep water should not change in response to sea level rise. However, for the same wind speeds and fetch lengths, waves generated over the continental shelf and shallower water will be higher and longer due to the reduced effects of bottom friction and steepness-limited breaking. From the Shore Protection Manual, U.S. Army Corps of Engineers (1984) the wave height H generated by wind speed U blowing over a fetch length F in water depth h is given by the expression

$$\frac{gH}{U^2} = 0.238 \tanh\left[0.53\left(\frac{gh}{U^2}\right)^{0.75}\right] \tanh\left\{\frac{0.025\left(\frac{gF}{U^2}\right)^{0.42}}{\tanh\left[0.53\left(\frac{gh}{U^2}\right)^{0.75}\right]}\right\} \tag{5.10}$$

In shallow water this reduces to

$$\frac{gH}{U^2} = 0.126 \left(\frac{gh}{U^2}\right)^{0.75} \tag{5.11}$$

It follows that

$$\frac{\partial H}{\partial h} = (0.126)(0.75) \left(\frac{gh}{U^2}\right)^{-0.25} = 0.75 \frac{H}{h} \tag{5.12}$$

and it is clear that wave height will increase with water depth. Wave period follows an expression similar to Eq. 5.10

$$\frac{gT}{2\pi U} = 1.20 \tanh\left[0.833 \left(\frac{gh}{U^2}\right)^{0.375}\right] \tanh \left\{ \frac{0.077 \left(\frac{gF}{U^2}\right)^{0.25}}{\tanh\left[0.833 \left(\frac{gh}{U^2}\right)^{0.375}\right]} \right\} \tag{5.13}$$

which in shallow water becomes

$$\frac{gT}{2\pi U} = \left(\frac{gh}{U^2}\right)^{0.375} \tag{5.14}$$

so that

$$\frac{\partial T}{\partial h} = \frac{(2\pi)(0.375)}{U} \left(\frac{gh}{U^2} \right)^{-0.625} = 0.375 \frac{T}{h} \tag{5.15}$$

and it is apparent that waves become longer as sea level rises.

After a wave leaves the storm area where it was generated, bottom friction will drain energy and reduce its height, but should not alter wave period. The losses due to friction can be expressed by the equation

$$\frac{\partial ECg}{\partial x} = - \overline{\tau_b u_b} \tag{5.16}$$

where E is energy density, Cg is group velocity, τ_b is the bottom shear stress and u_b the water particle velocity. The overbar denotes time-averaging over one wave period. Defining the instantaneous shear stress as

$$\tau_b = \rho \frac{f}{2} u_b |u_b| \tag{5.17}$$

performing the time average and integrating Eq. 5.16 yields

$$H(x) = \frac{H(o)}{1+\Gamma} \tag{5.18}$$

where

$$\Gamma = \frac{f\sigma^3 H(o) x}{3\pi g \, Cg \, \sinh^3 kh} \tag{5.19}$$

The effect of rising sea level will depend on the geometry of the continental shelf. For a uniform depth where the rise in sea level does not affect the shelf length, Γ will decrease and H at the shoreline will increase. However, if the geometry is such that a rise in sea level results in a wider shelf, Γ may tend to increase and H at the shoreline will then decrease, because bottom friction has had a longer time to drain energy.

Consider a storm on a shallow continental shelf that results in the conditions:

wind speed	U = 30 m/s ($g/U^2 = 0.01089$ m^{-1})
fetch length	F = 50 km
average water depth	h_o = 10 m

so that $gF/U^2 = 544.4$ and $gh/U^2 = 0.1089$. According to Eq. 5.10 waves will be generated whose heights are 2.06 m, and according to Eq. 5.13 the period will be 6.3 s.

Next consider the same storm after a 1 m rise in sea level. Following the same procedure as before yields H = 2.18 m and T = 6.4 s, or an increase in wave height and period of 5.8% and 1.6% respectively. If the continental shelf is 150 km wide and has a friction coefficient f = 0.01, the loss in wave height due to bottom friction is calculated using Eq. 5.18 and the wave height in the nearshore is found to be H = 0.82 m for the case without sea level rise. With the initial wave conditions for the 1 m sea level rise, the wave height on the inner shelf is found to be 0.96 m, or a 16.6% increase in wave height due to the combined effects of sea level rise during generation (slight) and reduced bottom friction on the shelf (marked).

More detailed numerical models for wind-wave generation have been developed, e.g., Cardone et al. (1976) and Resio (1981). Several models have been intercompared by the Sea Wave Modeling Project (SWAMP, 1985) but without definite conclusions due to lack of data. Cardone (1986) concludes that the level of error in wave height, period, and direction is on the order of 10% if high quality wind data are available. However, such data seldom are, and for predictive purposes the use of less accurate models for representing winds is often necessary.

5.4 RESEARCH NEEDS

From the previous analyses, it appears that long-term sea level rise will have a measureable effect on storm surge and wind wave generation only in locations where the continental shelf is shallow and its length fixed by a naturally hard shoreline, or one that has been stabilized with structures. Therefore, the aspects of storm surge and wind-wave generation that require research have less to do with long-term sea level rise, than with the basic phenomena themselves. Storm surge has received intensive theoretical and numerical study over the past three decades, and several sophisticated numerical models exist. However, there is a conspicuous lack of field measurements of hurricane and extratropical storm surge with which to calibrate and verify these models. Required are concurrent time series

from devices placed along the coast at intervals small enough to resolve the behavior of the surge as a storm moves out of the open ocean and makes landfall. The ability to model and predict storm surge cannot improve significantly without such data. Also, several phenomena associated with storms such as the superelevation of water level before arrival of the storm (often referred to as a forerunner) are still a mystery.

Research on wind-wave generation in deep and shallow water has progressed well. However, as noted there is a lack of detailed, high quality wind and wave data with which to verify these models. The basic process of damping of wind-waves as they cross the continental shelf due to bottom friction and breaking induced by wave-wave interaction are other areas in which research is needed. Theoretical work has progressed, but accurate field measurements are lacking. It is also necessary to stress the spectral approach to damping, as most methods available to date are limited to the assumption of monochromatic waves. Because waves refract in response to variations in water depth, basic research on the directionality of wave spectra, in both deep and shallow water, is also necessary before a better understanding of the effects of sea level rise on ocean waves can be assessed accurately.

Finally, if sea level is rising in response to changes in climate, these changes may (as spectulated by some) alter the frequency of occurrence, severity and behavior of hurricanes and storms. Research is needed not only to investigate these possibilities and improve predictive techniques, but the statistics and behavior from the historical record should continue to be examined so a reliable baseline can be established against which the future can be compared.

6. INTERACTION WITH NATURAL FEATURES AND CONSTRUCTED WORKS

6.1 INTRODUCTION

Assuming that sea level will rise a significant amount over the next century, and that shorelines will generally respond in some manner, the question arises as to by what means can (or will) this response be modified or prevented. Natural features such as shoals, headlands, inlets and even barrier islands themselves will cause the neighboring shorelines to respond in a manner different from that of the typical "open" coast. Man-made engineering works, e.g., breakwaters, jetties, and beach fills, by their very purpose alter shoreline response from that of nature, and so can modify shoreline response to sea level rise. Alternatively, the design, construction, and cost of coastal projects is highly dependent on local water depth. Relative sea level rise must therefore be addressed for a project having a long design life.

On a sandy coast, sea level rise generally invokes shoreline response by two mechanisms. First is simply the retreat due to flooding or inundation, which is often small because natural beach profiles are usually concave upwards in shape. However, the rise in sea level builds a large potential for additional erosion and shoreline retreat induced by wave action, which can be quite severe. The only means of preventing shoreline retreat due to inundation is by constructing dikes and seawalls. All other features which modify shoreline response, both natural and man-made, do so by altering or reducing the wave climate and have little effect on the inundation component. These features/structures are now discussed individually.

6.2 NATURAL FEATURES

Barrier islands - are the elongated natural islands composed of sandy material, which front a substantial portion of the mainlands of the world. These islands block out the wave activity to which the mainland shoreline would otherwise be subjected, essentially acting like large breakwaters. Although the mainland shorelines are still vulnerable to flooding due to sea level rise and wind-waves generated locally in the bays, barrier

islands are the paramount safeguard against realization of the full erosional potential of sea level rise in the back-bay region. This potential is especially strong because the sediments found here are often fine sands, silts, clay and peat, all highly erodible.

If sea level rise causes local barrier islands to deteriorate, breach more frequently, and "drown in place" rather than migrate landward, progressively more wave energy will penetrate through the chain and attack the mainland shoreline. This can result in enlargement of the bay area as the mainland erodes. An example is the Isles Dernieres on the coast of Louisiana, see Penland, et al. (1985). As shown in Fig. 6.1, since 1853 the large barrier has deteriorated drastically to become a series of five small islands, which have retreated about 2 km in 125 yr. Most of this is due to inundation and erosion accompanying the rapid subsidence of the delta region. Concurrently, Lake Pelto has been greatly enlarged by erosion of both the isles and the mainland. As the isles continue to disappear, erosion of the mainland should accelerate.

Shoals - are large deposits of sediment, usually associated with relict barrier islands, inlets, and large headlands. They serve to naturally limit the wave energy that impacts a shoreline, as a result of dissipation due to bottom friction and breaking, as well as partial reflection. As sea level rises, shoals become less effective unless their natural response is to grow, as is the case at inlets as described subsequently. Such growth of course requires sediment and may demand it from neighboring shorelines or inlets. An example of the effects of offshore shoals on the neighboring shoreline is Cape Canaveral, Florida, shown in Fig. 6.2. This cape has an extensive system of offshore and shore-connected shoals, which generally protect the cape from storm wave activity out of the northeast. Little protection is afforded from the southeast. The regional direction of net longshore drift is from north to south, and the offshore shoals A, B, C and the Hetzel Shoal have afforded enough protection for the Chester Shoal and False Cape to form utilizing this supply of sediment. The entire shoal system is responsible for the formation and protection of Cape Canaveral and Southeast Shoal. Farther to the south the shoreline assumes a crenulate shape as is common for such features. Field and Duane (1974) report that since 1878 Chester and

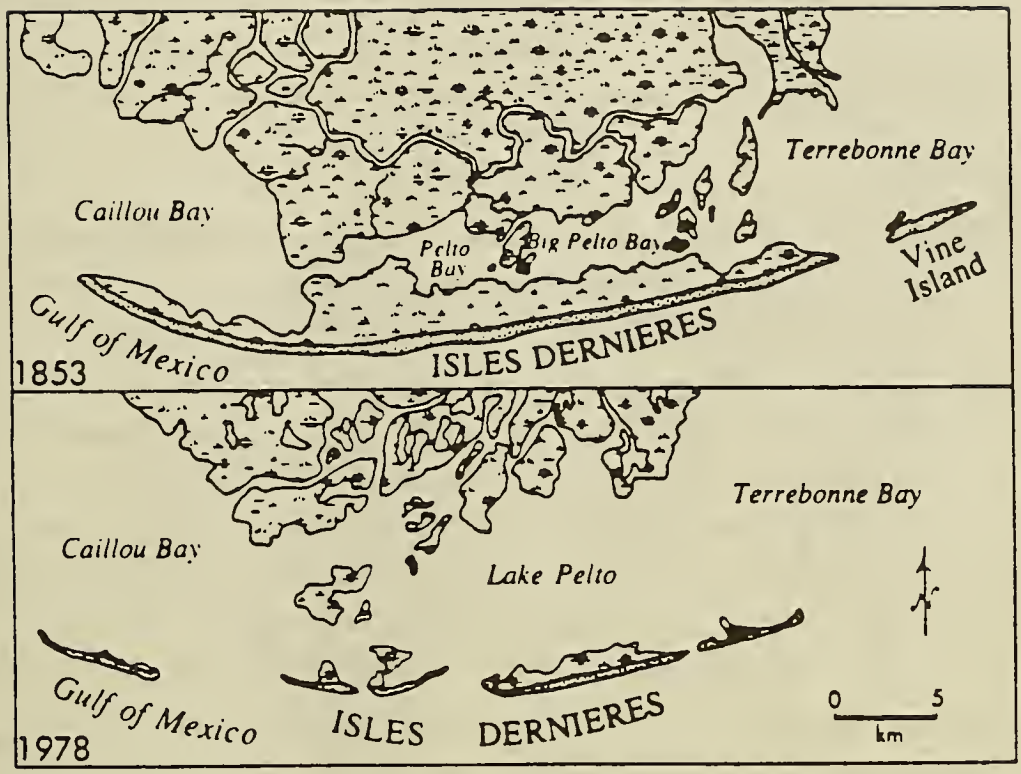


Fig. 6.1. Historical Shoreline Changes at the Isles Dernieres, Louisiana (after Penland et al., 1985).

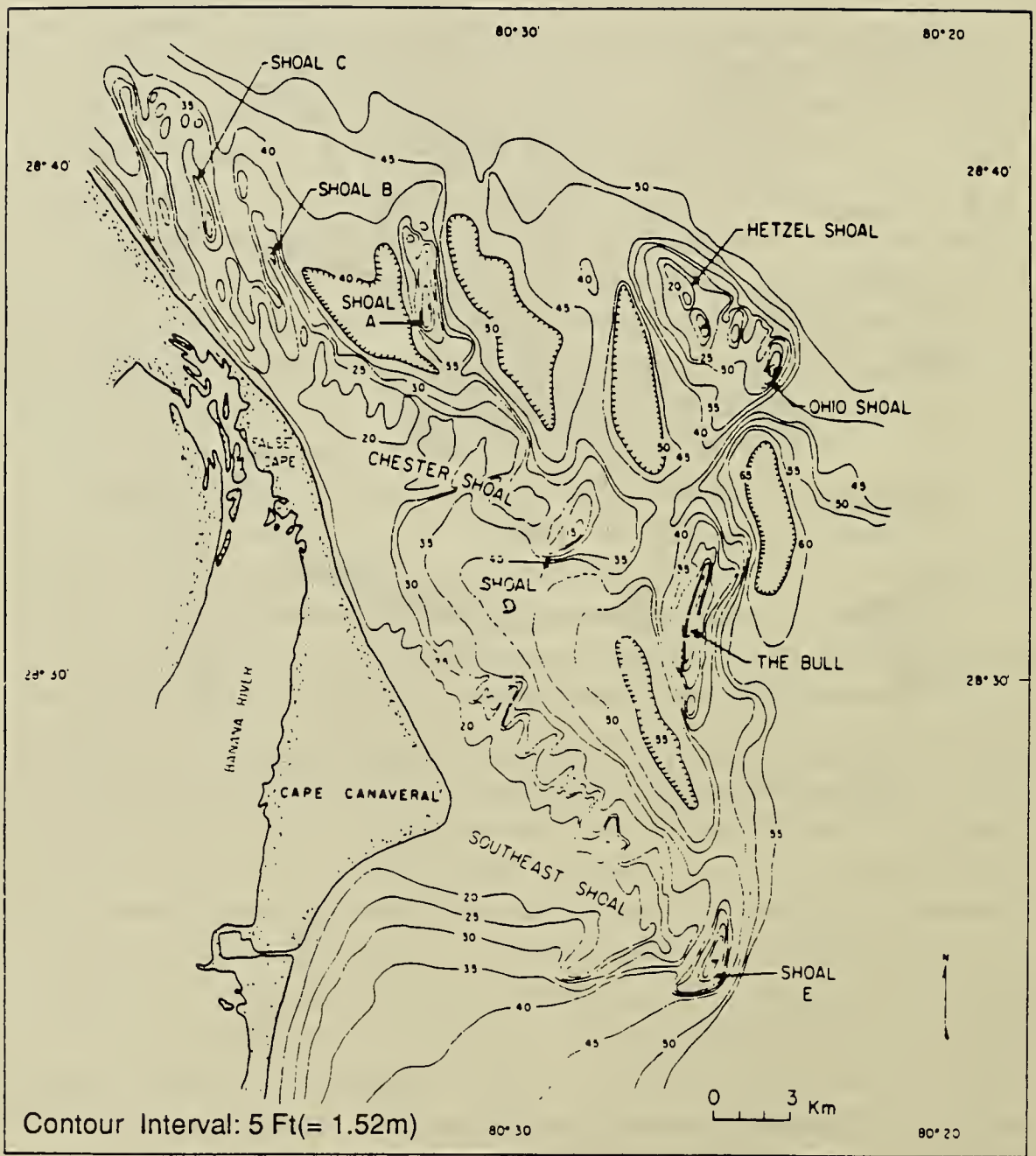


Fig. 6.2. The Shoal System at Cape Canaveral, Florida (after Field and Duane, 1974).

Southeast Shoals have become broader and thicker, and the offshore shoals have migrated slightly to the Southeast. Since 1898 accretion has occurred on the southern sides of Chester and Southeast, while the shoreline between these shoals and to the south of the cape has experienced erosion. This seems to indicate that as sea level has risen the protected areas at False Cape-Chester Shoal and Cape Canaveral-Southeast Shoal have continued to be maintained, apparently at the expense of the region to the south which has become more crenulate in shape due to blockage of longshore drift from the north.

Natural inlets - are the breaches between barrier islands, usually cut during storms. They generally affect neighboring shorelines as they migrate alongshore, which causes the updrift side to accrete and the downdrift side to erode. After a natural or man-made inlet is cut, ebb shoals along the mouth of the inlet grow and shunt sediment across the inlet and partially maintain the supply of sand to the downdrift beach. Sea level rise will tend to trigger a chain of events which could result in larger shoreline retreat than would occur if the inlet were not present. A rise in mean water level increases the depth and planform area of a bay and increases the hydraulic efficiency of an existing inlet, thereby increasing the tidal prism. An increase in tidal prism increases the velocities in an inlet which in turn may deepen or enlarge the throat. The ebb shoal then demands more sediment from the neighboring shorelines as the inlet grows in size. Sea level rise could possibly have a beneficial effect, in that dredging requirements may decrease. Circulation in bay and canal systems may also improve.

Natural tidal inlets also trap sediment in shoals on the bay side of the inlet (called flood shoals). Because of the reduced wave climate in the bay, these shoals are usually left behind as the inlet migrates and thus become a sink for sediment. As is the case with ebb shoals, the size of these shoals generally increases with the size of the inlet, and an increase in inlet size due to sea level rise will tend to remove more sediment from the beaches and store it in flood shoals.

A good example of the effect of long-term sea level rise on a natural inlet is Nassau Sound, Florida, shown in Fig. 6.3. Calculations of the

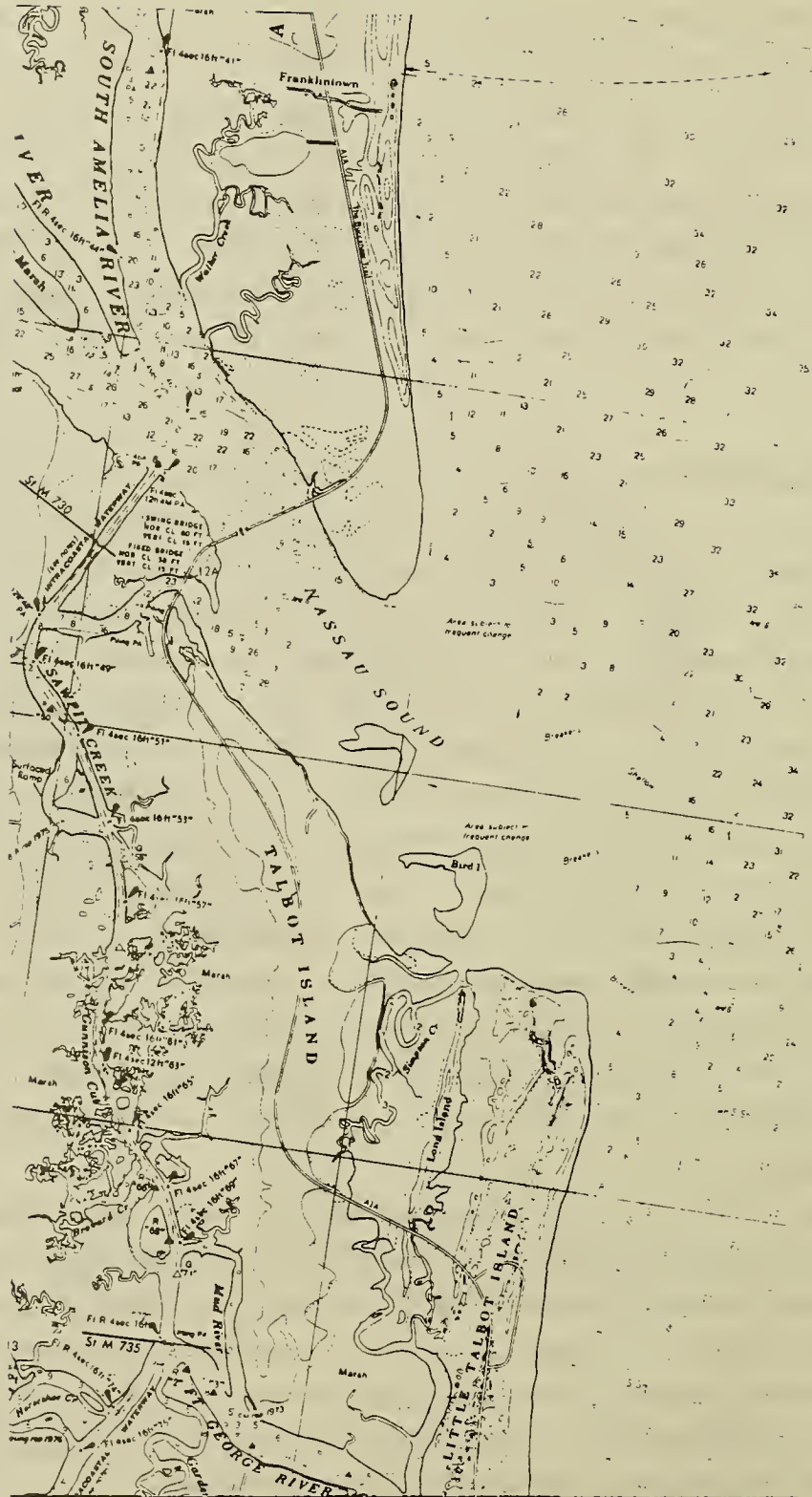


Fig. 6.3. Bathymetric Chart of Nassau Sound, Florida, Showing Ebb Shoals. Depths are in Feet (from NOS Nautical Chart 11489).

volume of sediment contained in the ebb shoals by Marino and Mehta (1986) indicate addition of 6.3 x 10^6 m^3 of material from 1871-1970. The shoal volume is currently about 40.5 x 10^6 m^3. During this 99 year span, relative sea level rose 0.3 m. Long-term sea level rise will also promote creation of additional inlets, which affects the behavior of existing ones. Each new inlet also has its own demand for sediment to maintain shoals.

Headlands - are natural intrusions of hard material on an otherwise sandy shoreline. These less-erodible features act as natural groins or breakwaters and compartmentalize a shoreline. A large single isolated headland usually causes a crenulate embayment to form on its downdrift shoreline, as is the case at Cape Canaveral. A series of two or more headlands spaced closely enough to act as a system will cause formation of embayments that are more semi-circular in shape.

The role that a headland or series of headlands will play in modifying shoreline response to sea level rise will depend on the amount of incident wave energy dissipated or reflected by the headland(s), and the aspect ratio (ratio of width to length) of the embayment(s). Those with broad faces parallel to the coast block significant amounts of energy, and their embayments have larger aspect ratios. This means they significantly increase the length of shoreline available to "resist" a given amount of wave energy - the amount being controlled by the fixed distance between headlands. Because of the reduced energy density at the shoreline, less of the potential erosion takes place as sea level rises. This situation is analogous to the performance of offshore breakwaters. However, narrow headlands do not block significant amounts of wave energy and although the shoreline in between may be reoriented, it is not lengthened substantially. Because the energy density at the shoreline is not reduced, little is done to affect on/offshore transport and therefore the full potential for erosion associated with a rise in sea level can be realized. This situation is analogous to a groin field, to be discussed subsequently along with offshore breakwaters.

An example of the effect of headlands on shoreline evolution is Wreck Bay on the west coast of Vancouver Island, British Columbia, shown in

Fig. 6.4. Quisitis Point and Wya Point are two natural headlands responsible for the large embayment in between. Although historical shoreline changes for the bay are not readily available, the general behavior in response to future sea level rise is expected to be as described. Part of the erosive potential of the rise will be spent on lengthening the shoreline of the bay as it enlarges. So, the average retreat of the shoreline will be less than on the open coast.

6.3 CONSTRUCTED WORKS

Dikes and levees - are free-standing, elongated mound-like structures used to prevent coastal and riverine flooding and to create usable land from low-lying, previously inundated wetlands. They are usually constructed of earth or sand (armored by clay, asphalt, rubble or vegetation), masonry, and concrete, and are often assisted by pumps to remove seepage. A few typical design cross-sections are shown in Fig. 6.5. Although costly to construct and maintain, dikes are the only means of totally preventing shoreline retreat (both inundation and wave-induced) due to a rise in relative sea level. They "modify" shoreline response essentially by creating a new shoreline at the structure location, and have been successful in many places throughout the world, more notably the Netherlands. If long-term sea level rise is significant, dikes may be the only workable means of protecting coastal cities.

The effect of sea level rise on existing dikes and the design of new ones is manifested predominantly in the required crest elevation of the structure. This is the elevation that prevents significant overtopping during the design storm. Crest elevation in turn determines the cross-sectional area of the structure and the volume requirements for material. A crude relationship is that the height increases directly with the rise in sea level and area increases with the square of the increase in sea level. More precise estimates depend on the actual design cross-section. The major sea level related question confronted in the design of new dikes and levees is whether to include projected long-term estimates or not. Answers depend on the site specific estimate of the rate of sea level rise, the type and method of construction to be used, the expected life span of the structure, and the expected frequency of

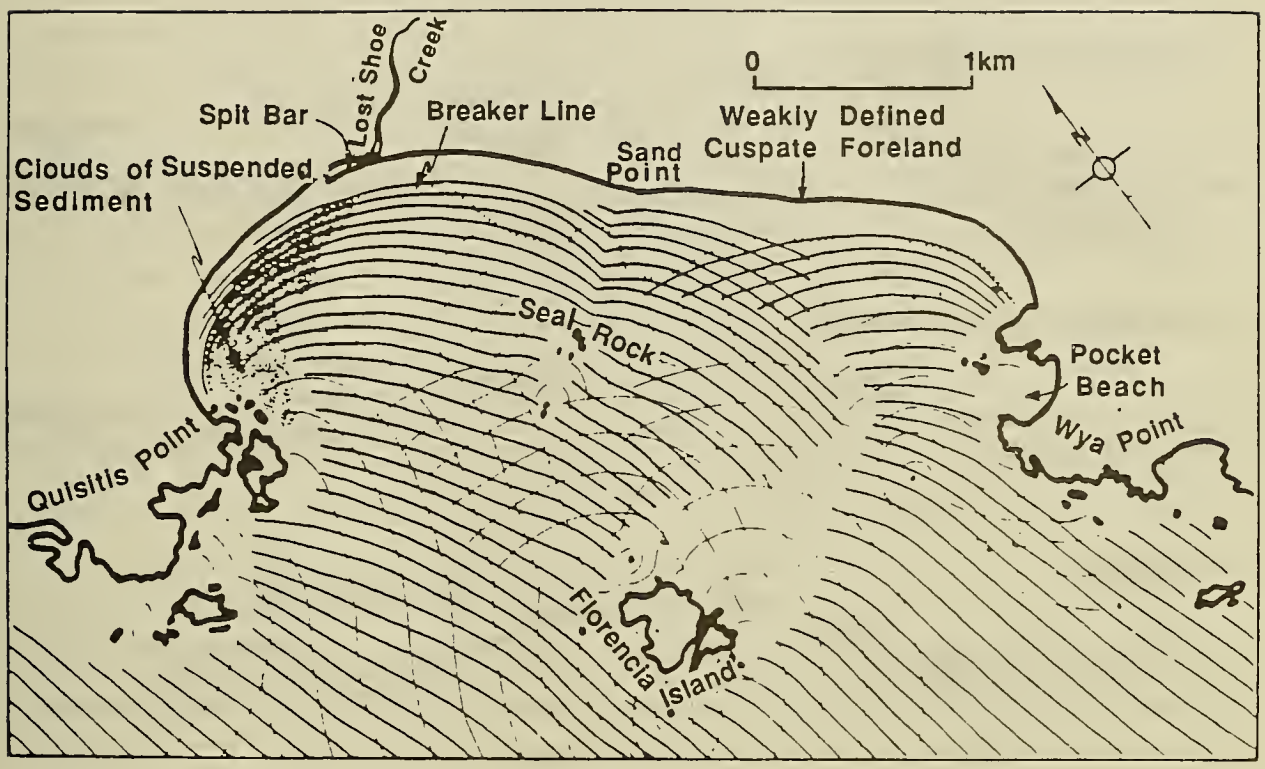


Fig. 6.4. Shoreline Between Two Headlands at Wreck Bay, Vancouver Island, with Observed Wave Patterns (from Bremmer and LeBlond, 1974, with permission from the Society of Economic Paleontologists and Mineralogists).

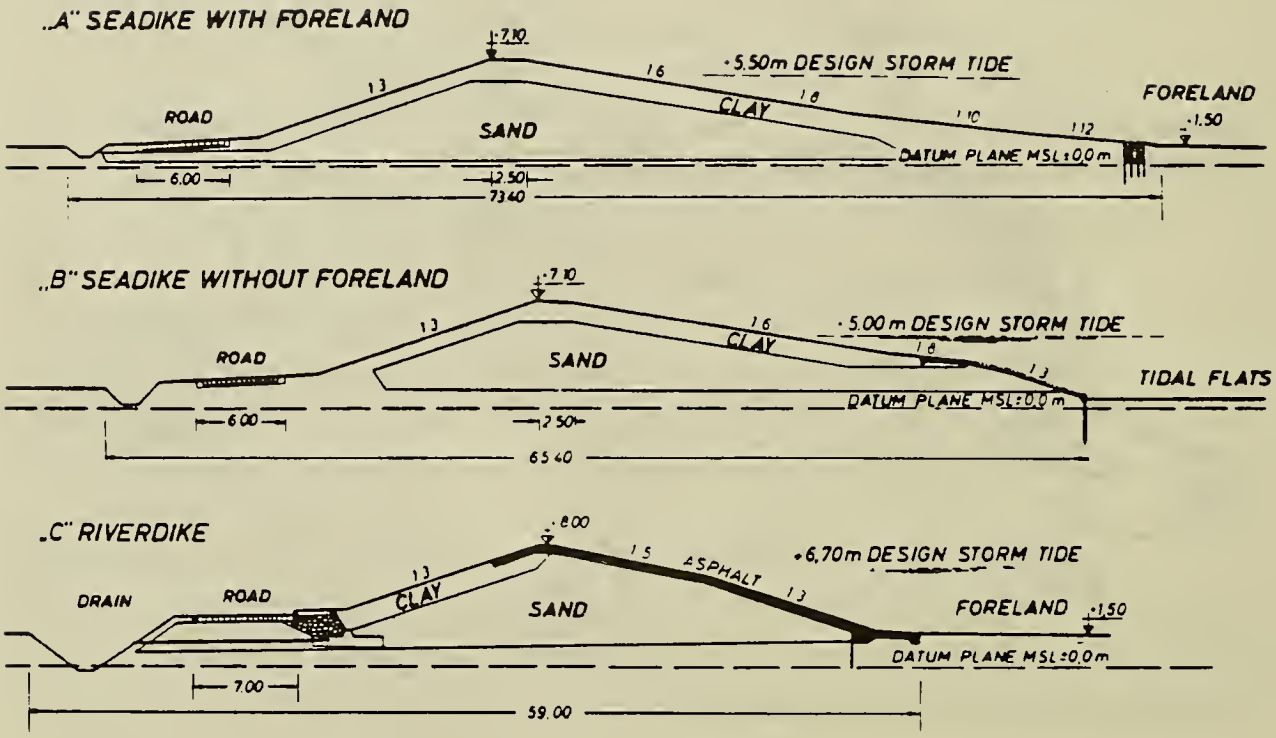


Fig. 6.5. Examples of Design Cross-sections for Sea Dikes (after Kramer, 1971).

maintenance. In regions where relative sea level is rising rapidly due to ground subsidence or tectonics, a projected estimate of suitable length (on the order of 10-50 years) may be most appropriate. "Hard" structures such as dikes built with masonry, concrete, and rubble are usually very expensive to maintain or improve, and should be designed using the maximum long-term sea level estimate projected during the design life of the structure. "Soft" structures built with sand or earth and armored with vegetation usually require frequent (but relatively inexpensive) maintenance and are more easily altered and improved. This permits raising of the crest elevation in response to actual sea level rise, rather than designing for a perhaps uncertain projection of sea level.

The best example of the use of dikes and levees to prevent coastal flooding, and their interaction with long-term sea level rise, is the Netherlands. Dikes have existed in the Netherlands since pre-Roman times and over 1,000 km now exist (Lingsma, 1966). However, several catastrophic failures during storms have served to periodically demand a review of their use and design, the most recent being the flood of February 1, 1953, where 1,783 people were killed and total damage was estimated at 250 million dollars (Wemelsfelder, 1953). This disaster prompted construction of the massive Delta Project, whose large storm surge barriers were just recently completed (see Kohl, 1986). As shown in Fig. 6.6, almost half of the Netherlands is below mean sea level and protected by dikes. The situation here is a clear microcosm of the future of many regions around the globe if the greatest estimates of sea level rise prove accurate.

Seawalls, bulkheads, and revetments - are structures of concrete, masonry, steel sheet pile, or rubble used to armor the shoreline and prevent retreat due to the combination of wave activity and sea level rise. Although performing much like dikes, they usually are not free-standing and are always "hard" features, with vertical or steeply sloping faces. They generally are used on a local rather than regional basis and are built to protect the upland long a limited section of beach. Besides their cost, the major drawback to seawalls is that as sea level rises progressively less sandy beach is available for recreation and additional storm protection. Periodic beach nourishment is often required as mitigation.

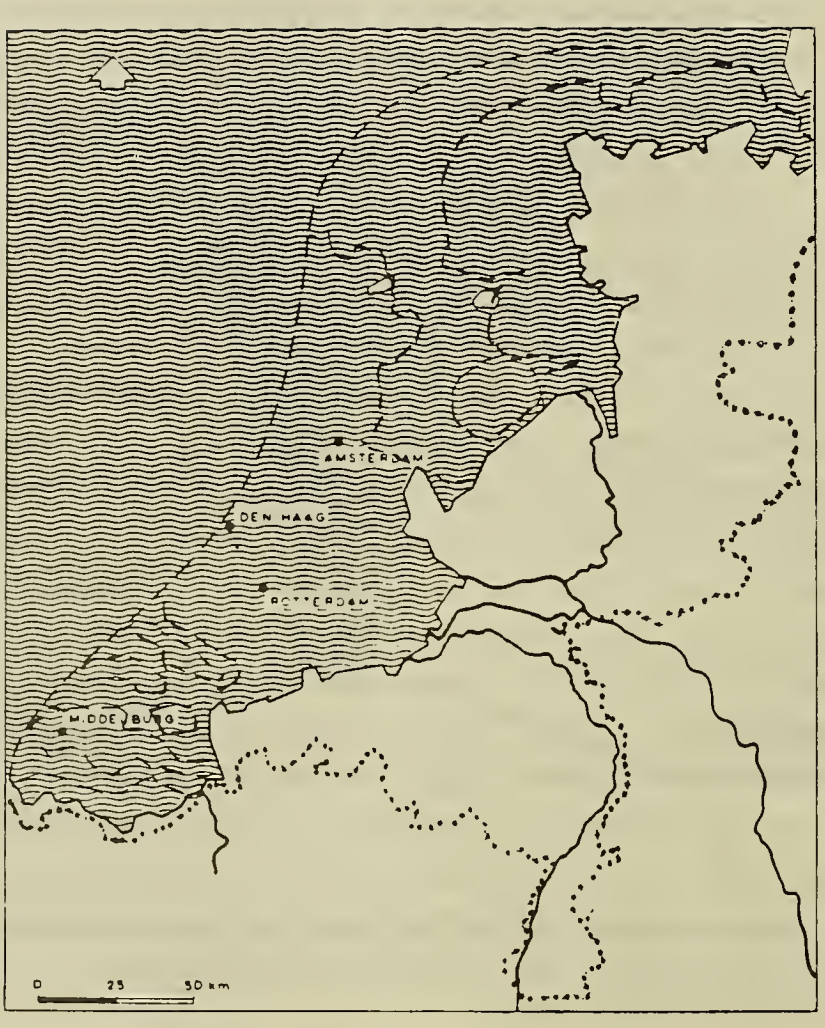


Fig. 6.6. Shoreline of Holland if There Were No Dikes, Showing a 50% Loss in Land Area (after Lingsma, 1966).

Typical cross-sections for a seawall, bulkhead and revetments are shown in Fig. 6.7.

Sea level rise affects the design, construction, and maintenance of these structures in the same general manner as with dikes. As sea level rises, higher crest elevations are required, but because the structures are not free-standing the required cross-sectional area increases more linearly than quadratically with sea level. As with dikes, a seawall, bulkhead or revetment can either be designed with enough crest elevation to account for projected sea level rise, or else the crest elevation can be periodically raised in response to sea level. Because these are hard structures, it is usually difficult and expensive to exercise the second option.
Galveston, Texas is fronted by a seawall constucted after the city was demolished during a major hurricane in 1900, in which more than 6,000 people were killed. The wall, whose cross-section was shown in Fig. 6.7a and planform is displayed in Fig. 6.8, is 4.9 m high and over 16 km long. Nine million cubic meters of fill were placed behind the wall and much of the city was raised in elevation. The seawall has been subjected to seven major storms since 1915, during which overtopping and toe scour have required additional fill and rubble toe protection. Subsidence of the wall has also been a problem, especially in places where it is located over a soft clay stratum. Relative sea level at Galveston has risen approximately 24 cm since 1904 (Leatherman, 1984), and during that time most of the original beach fronting the wall (up to 90 m wide) has been lost. Leatherman also indicates that diking will be necessary in the future to preserve the city.

Breakwaters - are free-standing structures, usually of rubble mound construction, attached to the shoreline or seaward (detached) of the shoreline. Breakwaters cannot prevent inundation by sea level rise, but can modify shoreline response by blocking some of the incident wave energy. The resulting shoreline (for detached breakwaters) has a bulge associated with each structure, and holds the mean shoreline at a more seaward position. Effective in preventing beach erosion due to both longshore and on/offshore transport, offshore breakwaters have been used for shore protection in the U.S., Canada, Europe, and quite extensively (over 2,500)

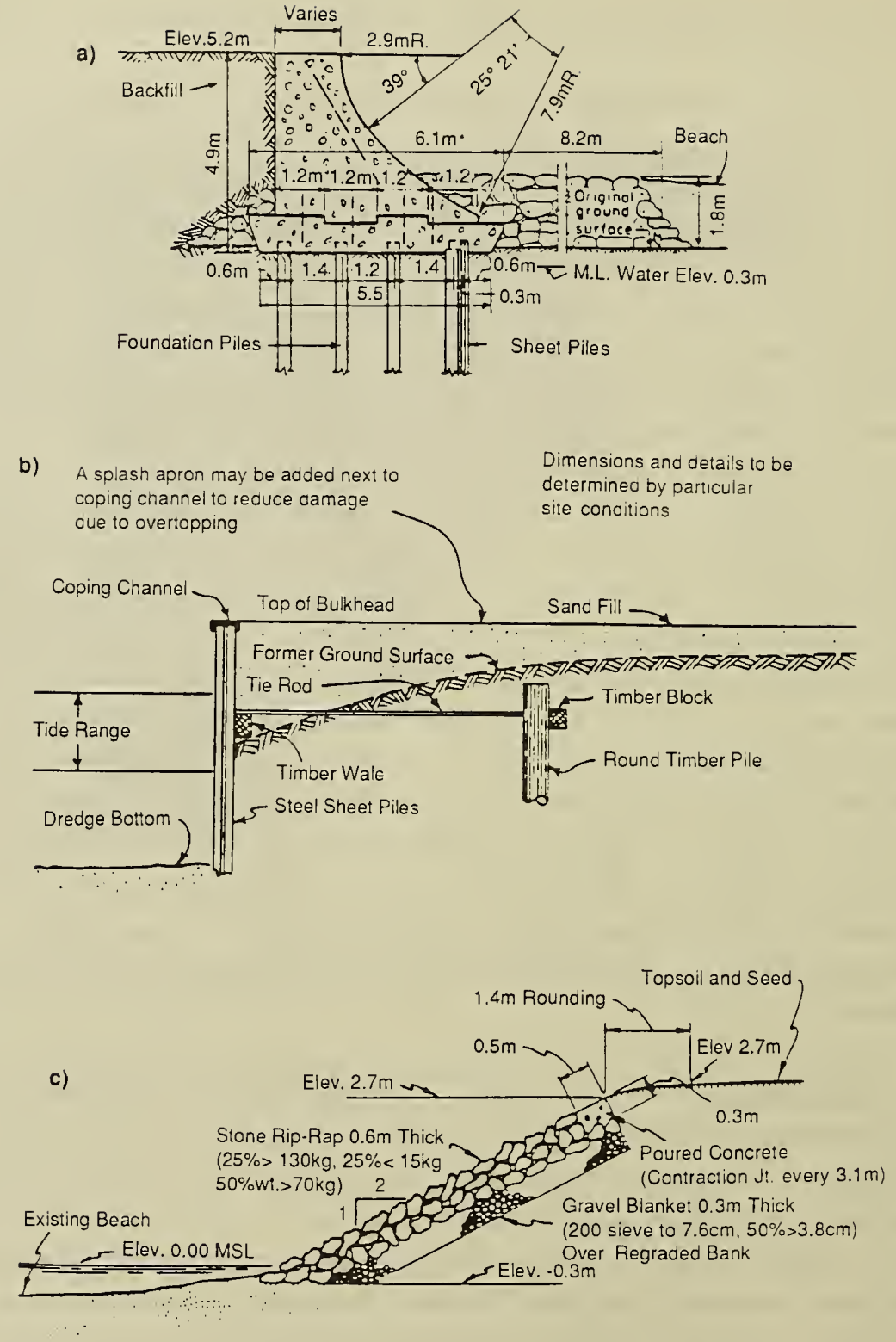


Fig. 6.7. Typical Cross sections of a) Seawall, b) Bulkhead and c) Revetment (adapted from Shore Protection Manual, U.S. Army Corps of Engineers, 1984).

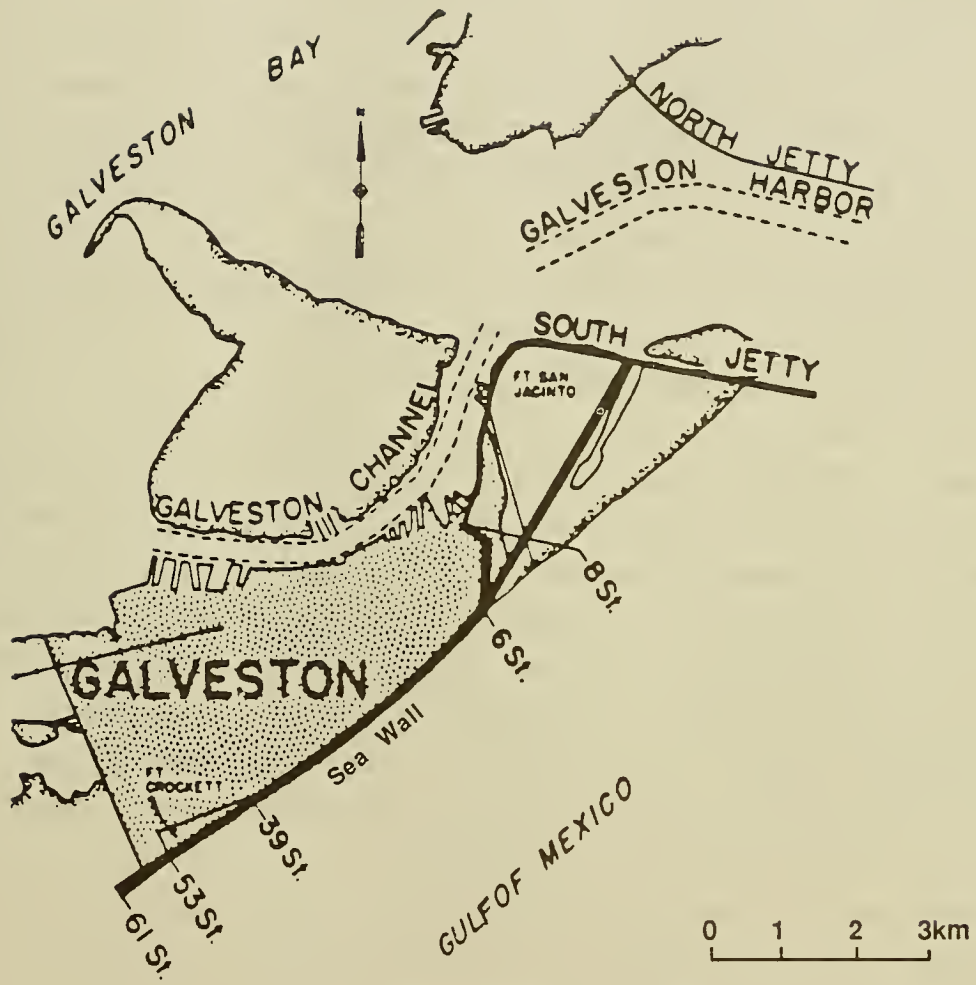


Fig. 6.8. Plan view of the Galveston Seawall (after Davis, 1952).

in Japan. Although initial construction costs can be high, proper design usually ensures low maintenance. The shoreline response and functional design of offshore breakwaters is extensively discussed in Dally and Pope (1986).

As sea level rises, an existing breakwater project will lose sediment from its salient(s) as its relative position moves offshore and overtopping becomes more frequent. In order to maintain shoreline position and a prescribed level of protection, the structure will need to be lengthened and its crest elevated. Otherwise, projected sea level must be used in both structural and functional design, with the margin of safety diminishing as sea level rises during the life of the project.

An example of a segmented breakwater project installed to provide shoreline protection and a recreational beach is found at Presque Isle, Pennsylvania. The project, shown in planform in Fig. 6.9, consists of three segments, each 38 m long and placed 46 m offshore of a beach fill. There is a substantial longshore drift (from left to right) from which the structures have entrapped additional sand to form a series of salients which progressively diminish in size in the drift direction. These salients erode during storms and accrete in calm weather, but the placed fill has remained relatively unscathed.

Groins - are shore-perpendicular structures made of timber, steel or concrete sheet pile, or rubble, whose purpose is to entrap sediment moving alongshore. The shoreline accretes on the updrift side and erodes on the downdrift until sand is able to pass around the end of the structure and restore the longshore drift. If the fillet is placed artifically during groin construction, much of the downdrift erosion can be prevented. For long stretches of beach, a groin "field" of many structures is used, examples of which are found at Rehoboth Beach, Delaware; West Hampton Beach, Long Island, New York; and Madeira Beach, Florida. It is stressed that groins are only useful if local erosion is due to spatial variation in the longshore drift, and have little positive impact if erosion is due to on/offshore sediment transport. Consequently, the use of groins will do little to modify shoreline response to sea level rise.

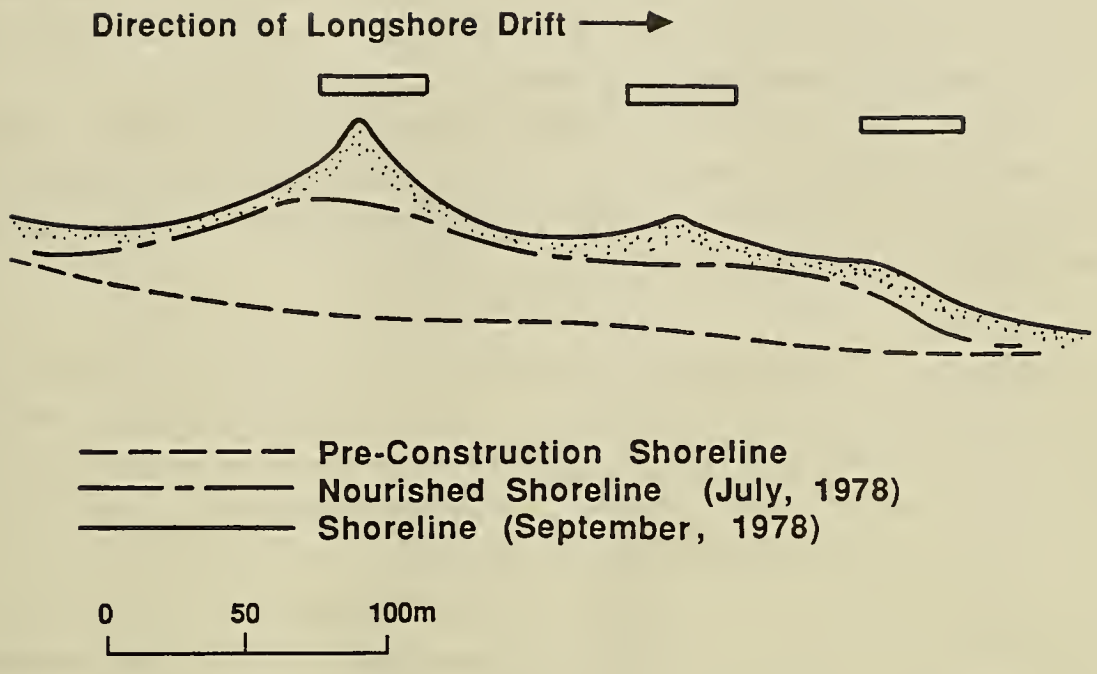


Fig. 6.9. Breakwater Project and Shoreline Response at Presque Isle, Pennsylvania.

Sea level rise will generally result in a loss of efficacy of existing groin projects. Increased water level will allow more overtopping by waves, and eventually flanking could occur at the landward end of the structure as the shoreline retreats. Groins with long useful lives may require lengthening and raising, while those with shorter life spans should be replaced with redesigned structures.

Figure 6.10 shows the groin field at Long Branch, New Jersey, where the longshore drift is from south to north. This project has succeeded in trapping sand and building a beach, but apparently at the expense of the North Long Branch shoreline.

Jetties - are shore-perpendicular structures, usually of rubble mound construction, placed at tidal inlets in order to stabilize their position and maintain a navigable channel. Shoreline response to construction of jetties is similar to that of groins, but on a larger scale as jetties are usually very long. A large fillet is formed on the updrift side of the inlet, with the downdrift shoreline often subject to severe erosion. Jetties serve to increase the velocities in a tidal inlet, which deepens the cross-section and pushes the ebb shoals offshore, entrapping even larger amounts of sediment. Without mechanical bypassing of sand from the updrift to downdrift side of an inlet, the downdrift beach will erode until the updrift fillet and ebb shoals are large enough to shunt sand across the inlet.

Jetties may exacerbate shoreline retreat as sea level rises in the following manner. A rise in mean water level increases the depth of a bay and increases the hydraulic efficiency of an inlet, thereby increasing the tidal prism. An increase in tidal prism increases the velocities through the inlet which in turn pushes the ebb shoal offshore into deeper water. The shoal then demands more sediment from the neighboring shorelines until the regional longshore transport rate is restored (if possible). As noted previously, long-term sea level rise will also promote creation of additional inlets, which if stabilized will each demand sediment to maintain ebb shoals.

As with groins, sea level rise will tend to reduce the efficacy of jetties due to overtopping and possible breaching at the shoreward end.

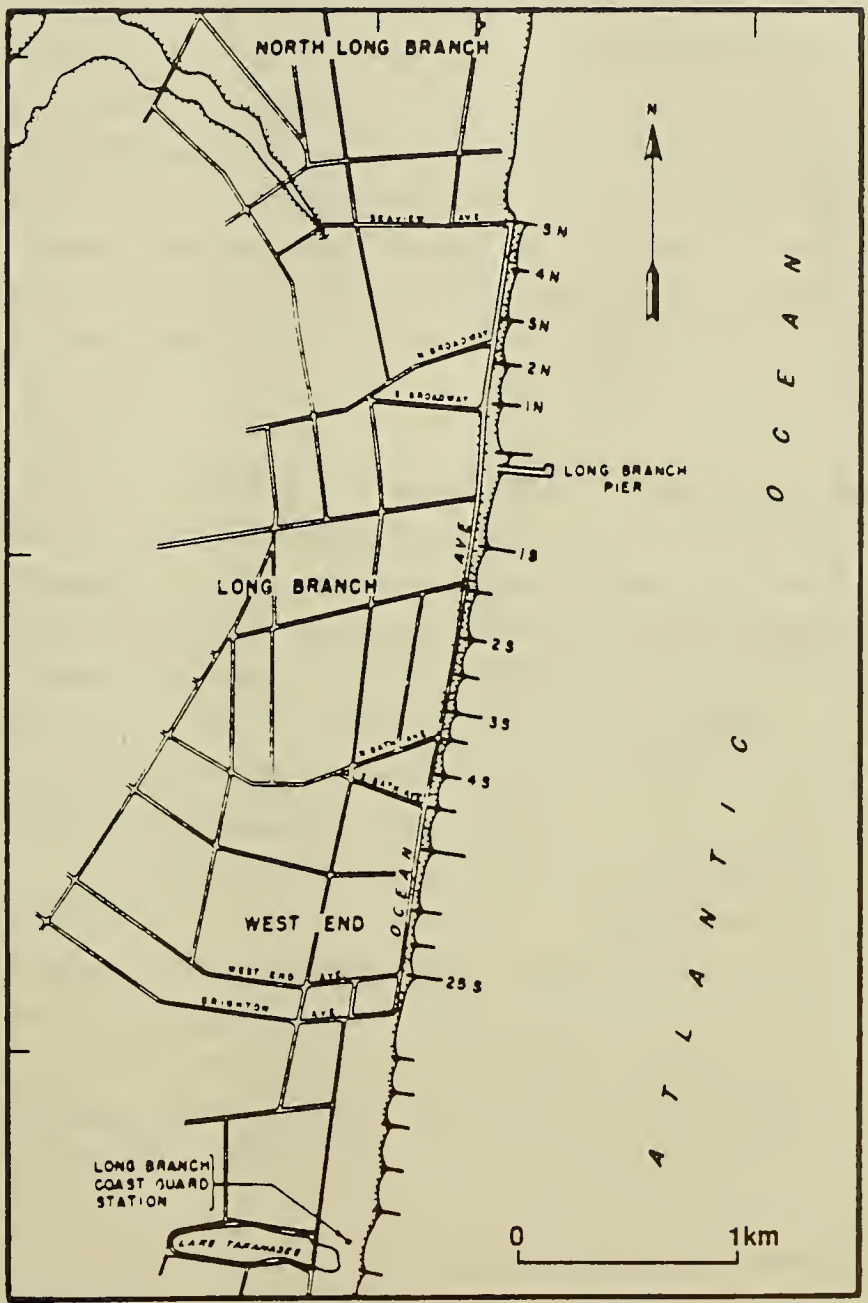


Fig. 6.10. Groin Field at Long Branch, New Jersey (after Army Corps of Engineers, 1964).

Because jetties usually have long life spans, they may require lengthening and raising of the crest of the structure. The possible increase in tidal prism and current magnitudes could increase scour and undermine jetties. The jetties at Indian River Inlet, Delaware are currently threatened by scour induced by an increase in tidal prism.

A typical example of the shoreline response to jetties is the inlet at Ocean City, Maryland shown in Fig. 6.11. A hurricane cut the inlet in August, 1933 and jetties were constructed shortly thereafter. By 1976 the updrift shoreline had advanced 245 m while the downdrift had retreated 335 m. The shoreward end of the south jetty has had to be rebuilt and extended several times. It is doubtful that in the 50 years since "stabilization," the shorelines have regained a state of dynamic equilibrium, so the effects of 50 years of sea level rise cannot be deduced accurately.

Beach nourishment - is the mechanical placement of sand on a beach to advance the shoreline. It is a "soft" protective and remedial measure that leaves a beach in a more natural state than hard structures, and preserves its recreational value. Beach fills cannot "modify" shoreline response to sea level rise because the natural littoral processes remain unaltered, and thus fills can only be regarded as a temporary measure. Although requiring maintenance at regular intervals and after severe storms, beach fills have been successful in many instances such as Miami Beach, Florida; Virginia Beach, Virginia; and Wrightsville Beach, North Carolina.

The greatest effect of long-term sea level rise on beach fill design is to increase the volumetric requirements of the fill and so increase costs. Attempting to hold the shoreline in one location will necessarily require a steeper beach profile as sea level rises. This means increased volumes of placed sand are necessary to satisfy the offshore transport demand, or else placing material of coarser grain size than the native sediment.

An example of a successful beach nourishment project is that at Harrison County, Mississippi, shown in Fig. 6.12. Constructed in 1951-1952 of 4.6 million m^3 of fill, the project provided 280 hectares of new beach which was 90 m wide with a berm height of 1.5 m, and fronted the seawall

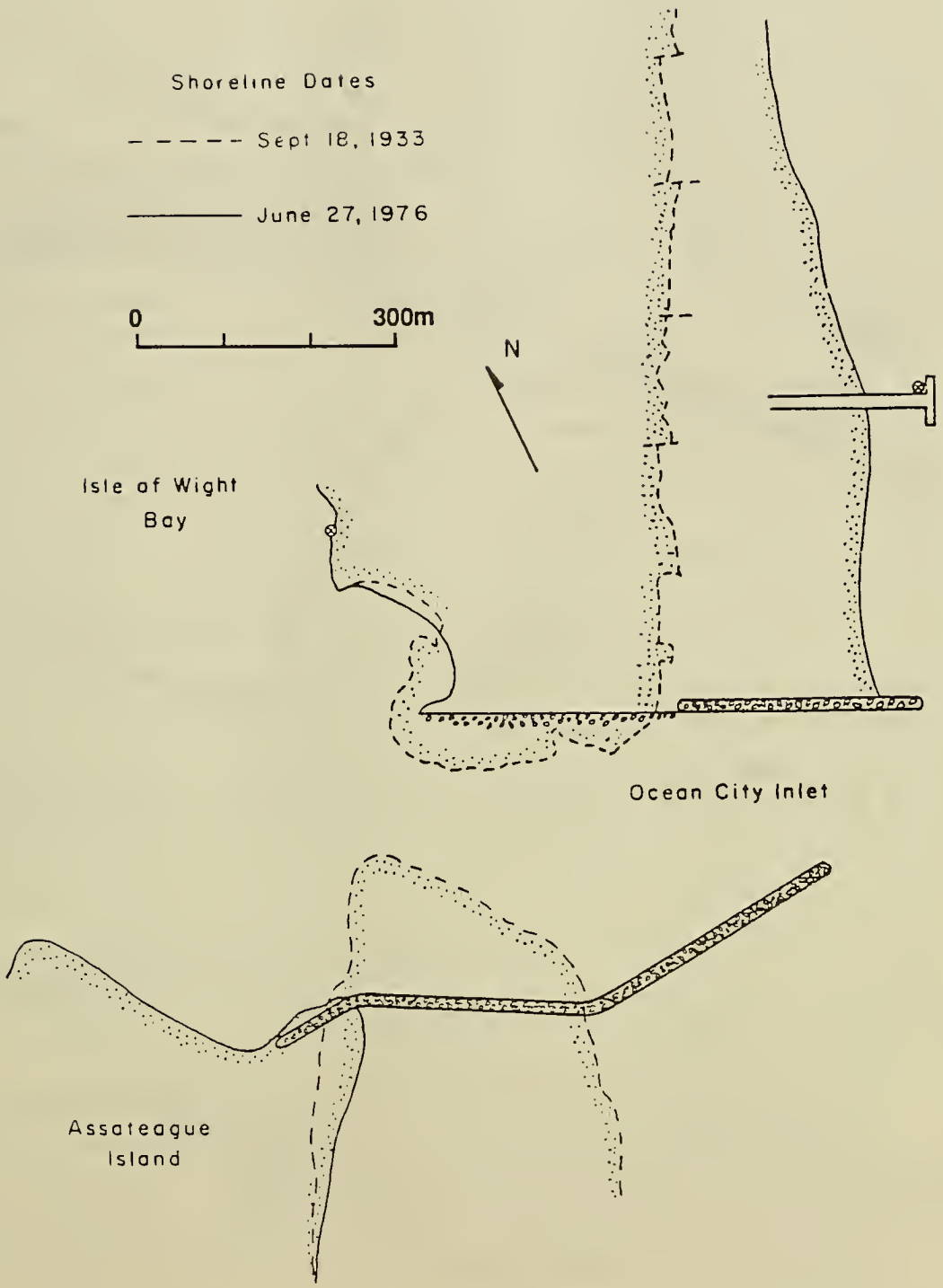
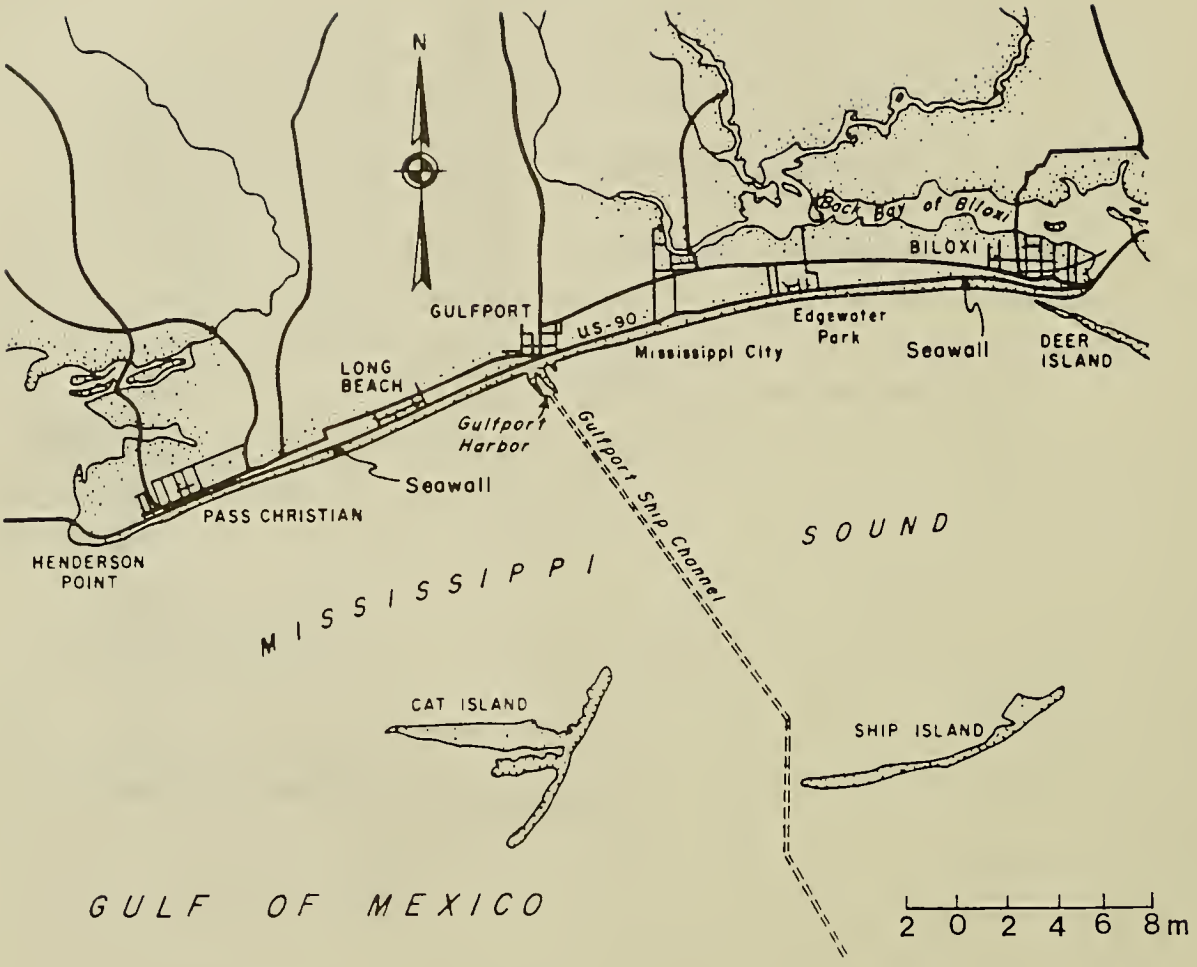


Fig. 6.11. Shoreline Response to Jetty Construction at Ocean City, Maryland (after Dean et al., 1979).



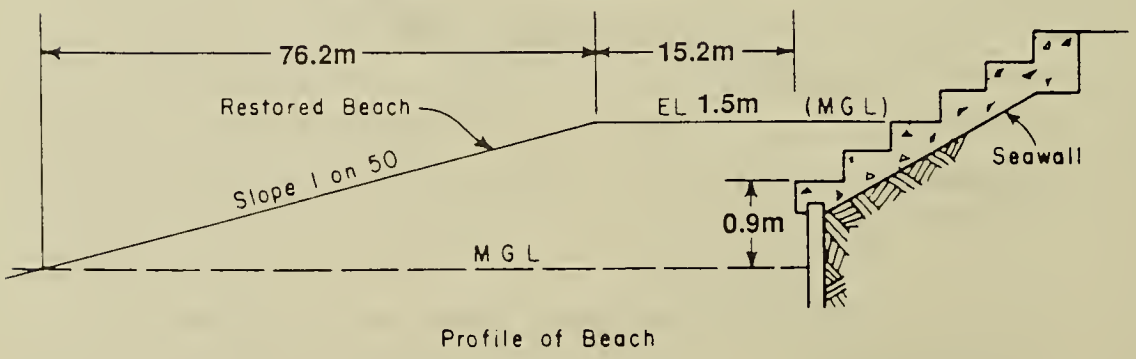


Fig. 6.12. Beach Nourishment Project at Harrison County, Mississippi (after Army Corps of Engineers, 1984).

constructed in 1925-1928. The project has performed well, with annual losses on the order of 76,500 m^3, and has provided upland protection during several major hurricanes. Several islands provide some shelter to the project from the Gulf of Mexico and may be partially responsible for its longevity. It was renourished with 1.5 million m^3 of fill in 1972-1973, following the effects of hurricane Camille (1969), which caused storm tides locally in excess of 6 m. Relative sea level is estimated to have risen only 8 cm during the life of the project (Hicks et al., 1983), forestalling conclusions of the fill's stability in response to sea level rise.

6.4 COST OF COASTAL WORKS

Although the effect of a rise in relative sea level on the cost of a coastal structure or beach nourishment project can only be accurately determined on a case-by-case basis, several crude indicators are available. For rubble mound structures, the cost increases with the required individual weight of the armor stone. Using the well known Hudson formula, found in the Shore Protection Manual (Army Corps of Engineers, 1984), the weight (W) increases with the cube of wave height. From section 5, expression 5.10 for the generated wave height and 5.18 for the height after bottom friction can be used to determine the relative increase in stone weight. For the example presented (sea level rise of 1 m, wind speed of 30 m/s, fetch length of 50 km and shelf depth of 10 m), the ratio of weights is

$$\frac{W \text{ (after s.l.r.)}}{W \text{ (before s.l.r.)}} = \frac{(0.96)^3}{(0.82)^3} = 1.60 \qquad (6.1)$$

or a 60% increase in stone size. We see that sea level rise may have a significant impact on the design and cost of rubble mound structures. Also, the future cost of buying additional land required to raise the crest of a levee or dike often is much greater than the additional material required, and should be considered in long-range planning and design.

For beach nourishment projects, the increase in the rate of losses can be examined by assuming the transport rate (Q_S) to be proportional to wave

height to the 2.5 power (Dean, 1976). For the same example of section 5 reiterated above, this means that

$$\frac{Q_s \text{ (after)}}{Q_s \text{ (before)}} = \frac{(0.96)^{2.5}}{(0.82)^{2.5}} = 1.48 \qquad (6.2)$$

or a 50% increase in the rate of losses from a beach fill.

Using two different methods, the approximate cost of maintaining the existing shoreline of Florida with beach nourishment was examined in the NRC report "Responding to Changes in Sea Level: Engineering Implications" (National Research Council, 1987). For the three different scenarios of sea level rise examined, the annual costs range from $33 to $204 per linear meter of shoreline, or between 0.1 and 3.4% of the present day value of beachfront property in Florida. The amount varied by a factor of 2.5 between the two methods - highlighting the need for research in this area.

6.5 RESEARCH NEEDS

Research needs in the area of modifying shoreline response to sea level rise and the effects of sea level rise on the design of protective works lie in the realm of ongoing basic studies of natural shoreline processes, and have little requirement for specific treatment of sea level rise. If engineers had a surf zone sediment transport model capable of reproducing and predicting beach response to storms and structures, including the effects of sea level rise would mean an almost trivial matter of increasing the mean water depth in the model. However, until the knowledge of basic processes has grown considerably and such models are developed, there is little reason to expect accurate prediction of the response of beaches to sea level rise to be possible, and that cost-effective techniques for modifying the response will be available.

There are four major areas requiring research in basic physical processes: 1) wave refraction/diffraction, 2) wave breaking, 3) undertow and longshore currents (nearshore circulation) and 4) sediment entrainment under shoaling and breaking waves. The knowledge gained from research in these areas would then be used as input to beach profile and planform response models.

Once a reasonable expertise in shoreline modeling has been reached, the greatest research need is for the engineering community to analyze and quantify the performance and costs of the available alternatives for dealing with sea level rise, and to then determine their cost-effectiveness. Studies should be implemented that are specifically devoted to dikes and artificial dunes, offshore breakwaters, and beach fill design. These measures appear to be the most promising for confronting sea level rise.

Research needs in structural design also require studies of a basic nature to increase performance and decrease costs. Most importantly, the most likely sea level rise scenarios must be refined and agreed upon before engineers will be inclined to include them in their designs. Until then, research on economical methods to retrofit existing structures must be pursued. The modes and methods of failure of coastal and estuarine structures, as well as self-induced subsidence must also be studied.

7. SHORELINE RESPONSE MODELING

7.1 INTRODUCTION

A potential dominant effect of relative sea level rise is shoreline erosion. An erosional trend on a developed coastline always requires a decision to: 1) retreat, 2) stabilize through coastal structures, or 3) stabilize through nourishment. Each of the above can be costly; accepting that under a given scenario of relative shoreline stability, sea level rise, etc. there is an "optimal" choice, it follows that an inappropriate choice could be inordinantly expensive. Given that eustatic sea level rise affects shorelines on a global basis, that the human rate of shoreline development is increasing and that some projections of future sea level rise are much greater than in the past, it becomes important to attempt to predict the shoreline response to such a rise.

Shoreline response to sea level change depends not only on the rate of change, but also on antecedent conditions and the degree and type of disequilibrium of the shoreface. The dominant engineering approach to predicting shoreline response is the so-called "Bruun Rule" which considers only cross-shore conditions and an offshore "closure depth" seaward of which there is no sediment exchange. The Bruun Rule yields a simple relationship resulting in horizontal shoreline retreat of approximately 50-100 times the rise of sea level. This chapter presents a more complete consideration of the sediment budget on the shoreface and attempts to remove some of the limitations of the Bruun Rule. Specific cross-shore components not included by Bruun but which could be of significance are: 1) shoreward transport of sediment across the shoreface, 2) deposition of suspended sediment, and 3) biogenic production of sediment. An important factor relating to shoreward sediment transport is the history of sea level change over the past ~20,000 years, with the last 6,000 years or so representing a relative still stand.

7.2 LITERATURE REVIEW

Prior to discussing the models for shoreline response, it is instructive to review estimates of sea level rise over the last

20,000 years or so, shown in Fig. 7.1. Sea level rose rapidly (about 0.8 m/century) from 20,000 years before present (BP) to about 6,000 years BP. Over the last 6,000 years, sea level has risen at the greatly reduced rate of 0.08 m/century, which is roughly consistent with estimates of 0.11 m/century based on tide gage data over the last century. As will be discussed later, the earlier much more rapid rise of sea level may still be having an effect.

The most widely applied engineering approach to predicting shoreline response to sea level rise is the so-called Bruun Rule. This rule considers: a) the active profile to always be in equilibrium, and to retain its relative position to sea level, and b) the active portion of the profile to be limited by the "depth of effective motion" seaward of which no sediment exchange occurs. With the above assumptions, when sea level rises a vertical distance, S, the entire active profile must rise also by S, requiring a volume $\forall_R$, of sand per unit beach length

$$\Delta\forall_R = SL \tag{7.1}$$

in which L is the offshore length of active profile. This required sand is provided by a profile retreat, R, over a vertical distance, h_*+B (see Fig. 7.2). The volume generated by this retreat is

$$\Delta\forall_+ = (h_* + B)R \tag{7.2}$$

and equating the two volumes, the retreat R can be shown to be

$$R = S\,\frac{L}{h_* + B} = \frac{S}{\tan\theta} \tag{7.3}$$

in which θ is the average slope of the active profile out to its limit of active motion, Fig. 7.3. From Eq. 7.3, it is clear that beach profiles with mild slopes would experience greater recessions due to a given sea level rise than would steeply sloping profiles.

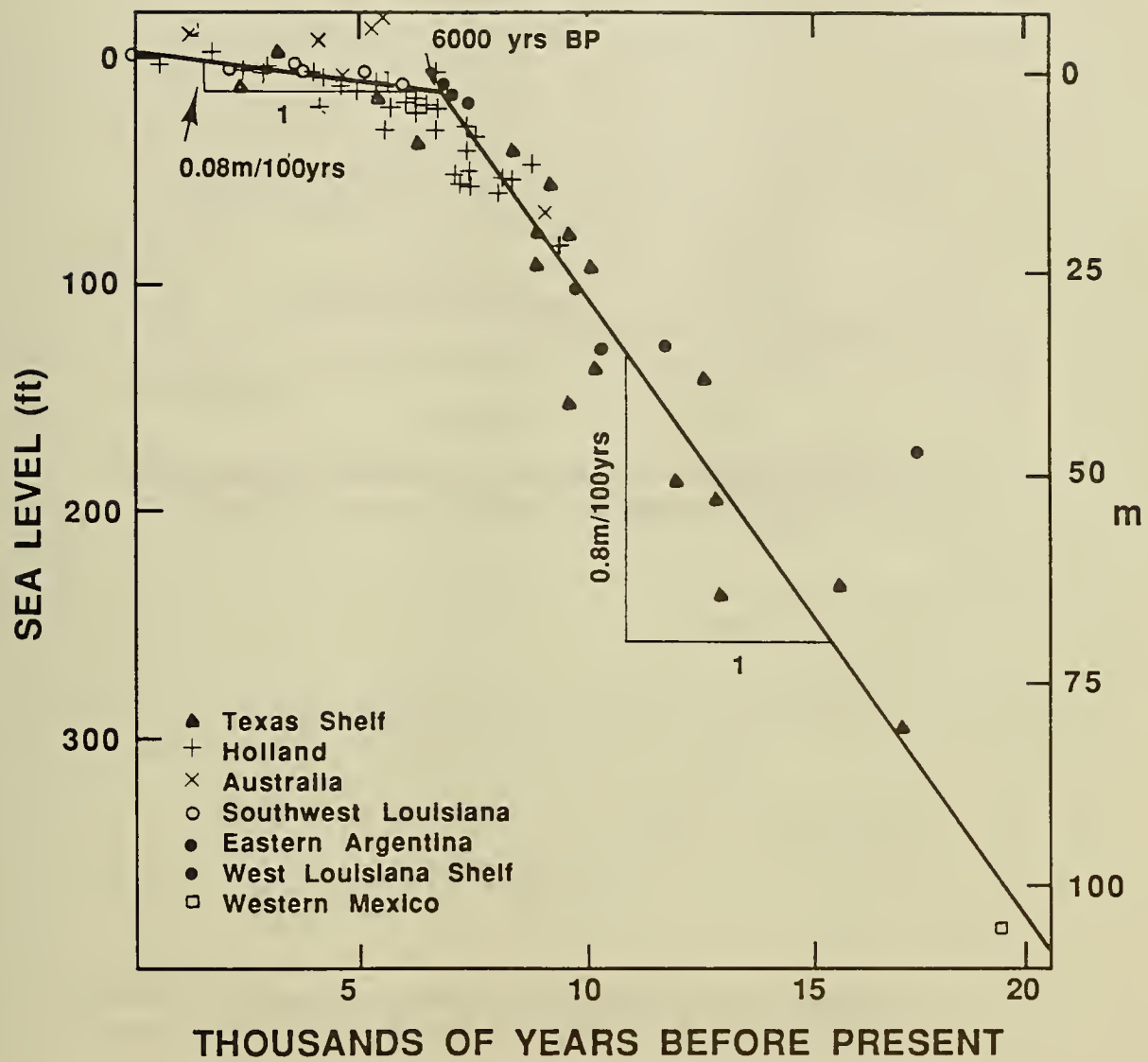
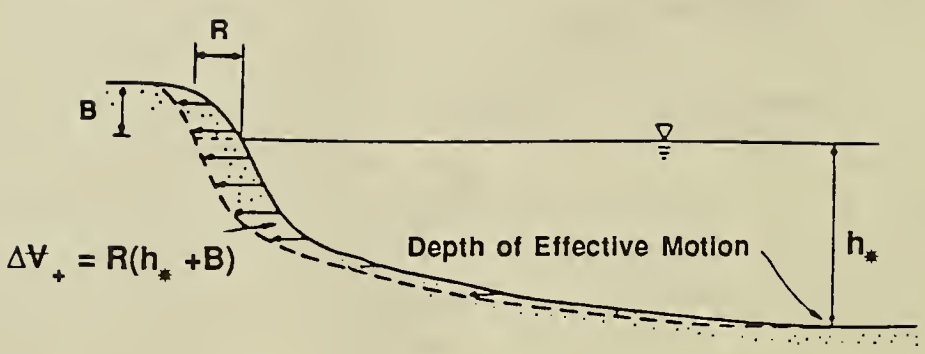
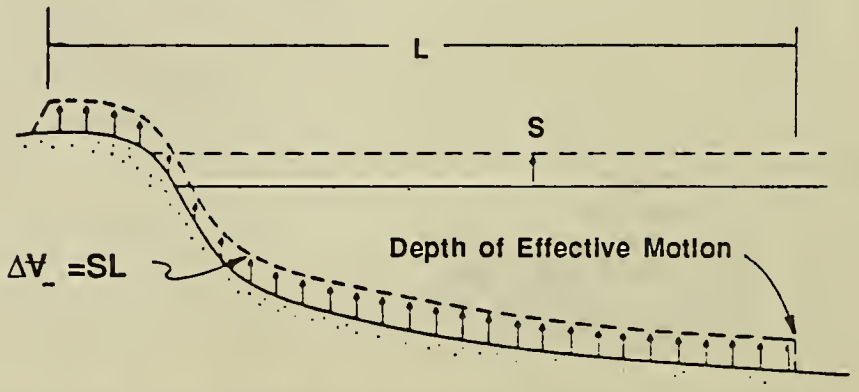


Fig. 7.1. The Rise of Sea Level as Obtained from Carbon-14 Dates in Relatively Stable Areas (from Shepard, 1963). Break in Slope some 6000 Years BP may have Provided Basis for Barrier Island Stability.



a) Volume of Sand "Generated" by Horizontal Retreat, R, of Equilibrium Profile Over Vertical Distance ($h_* + B$)



b) Volume of Sand Required to Maintain an Equilibrium Profile of Active Width, L, Due to a Rise, S, in Mean Water Level

Fig. 7.2. Components of Sand Volume Balance Due to Sea Level Rise and Associated Profile Retreat According to Bruun Rule.

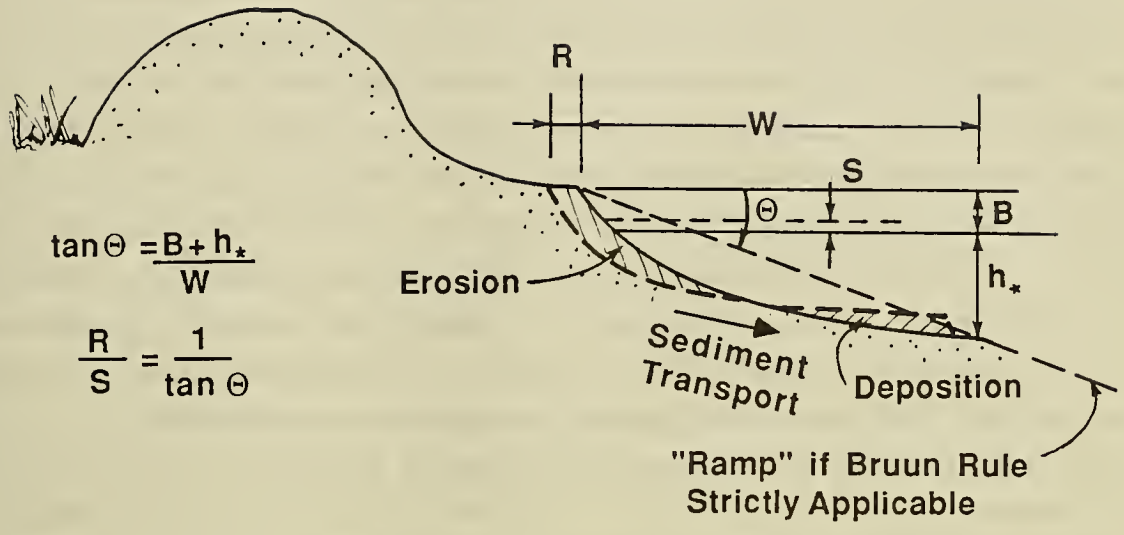


Fig. 7.3. The Bruun Rule with Only Seaward Transport of Sediment and Trailing Ramp Seaward of Active Profile.

Several laboratory and field studies have been carried out to evaluate the Bruun Rule, usually with confirmation claimed. Schwartz (1965) conducted small-scale laboratory model studies to determine whether an increase in water level caused an offshore deposition equal to the rise in water level as predicted by the Bruun Rule. The wave basin was quite small using medium sized sand of 0.2 mm. Following the development of an equilibrium profile, the water level was increased by 1.0 cm and the test program resumed with the same wave conditions. Following profile equilibration, it was found that the offshore profile had increased in elevation by 0.9 cm which Schwartz considered as confirmation of the Bruun Rule. Schwartz (1967) also conducted a second series of tests with slightly larger facilities, but following the same general experimental framework. Again good agreement with the Bruun Rule was reported. Field measurements were also carried out by Schwartz at Cape Cod, MA in which the shoreline response between spring and neap tides was evaluated in terms of the Bruun Rule. Although "a recognizable upward and landward translation of the profile was noted in the interval between neap and spring tides" was reported and the results were generally regarded as confirmatory, examination of the results is not convincing as to their significance to and agreement with the Bruun Rule. Also, it is not clear that spring tides, which of course have water levels both higher and lower than the average, should be equated to a sea level rise since the average water level is unchanged. Moreover, it is not clear that the Bruun Rule was meant to apply on such a short-term basis especially recognizing that short-term changes in wave climate and convergences of longshore sediment transport can play an important role in beach profile changes.

Dubois (1975, 1976, 1977) has reported on shoreline changes in Lake Michigan in association with a 30 cm rise over a 35 week period. The shoreline recession of 7 m was regarded as substantiation of the Bruun Rule.

Rosen (1978) has evaluated the Bruun Rule on the Virginia shoreline of Chesapeake Bay. Using 14 beach profiles, Rosen found that the errors in predicted erosion rates on the eastern and western shores were +58% and -7% with the positive percentages indicating that the predicted erosion exceeds

the measured. As expected, considering smaller groups of profiles, the errors were larger.

Hands (1983) has evaluated the Bruun Rule employing a series of 25 profiles along 50 km of the Lake Michigan eastern shore over a 7-year period. During this period, the water level rose by 0.51 m and then fell by 0.31 m. Fig. 7.4 from Hands shows that the shoreline responded to the changes in water level, although with a lag. Hands recommends that in the absence of other information the "depth of limiting motion" be taken as twice the significant wave height.

Everts (1985) presented a sediment budget approach which encompassed and extended beyond the Bruun Rule. The method was applied to Smith Island, Virginia, and a 75 km segment of the Outer Banks of North Carolina to determine the portion of the shoreline retreat explainable by sea level rise. It was found that 55% and 88% of the measured shoreline retreat was attributable to sea level rise at Smith Island and the Outer Banks, respectively. The remaining component was interpreted to be due to gradients in longshore sediment transport. The sediment budget approach applied by Everts recognizes the limitations of the Bruun Rule and the need to consider a more complete framework for representing and interpreting shoreline response to sea level rise.

Dean and Maurmeyer (1983) have generalized the Bruun Rule to barrier island systems that retreat as a unit filling in on the bay side to maintain their width as they erode on the ocean side. Employing the notation of Fig. 7.5, the shoreline recession, R, due to a sea level rise, S, is

$$R = S \frac{(L_L + W + L_o)}{h_{b_o} - h_{b_L}} \tag{7.4}$$

It is clear from Eq. 7.4 that the recession will always predict a greater erosion than the Bruun Rule because: 1) a greater horizontal dimension is being elevated with sea level rise (the entire active barrier island

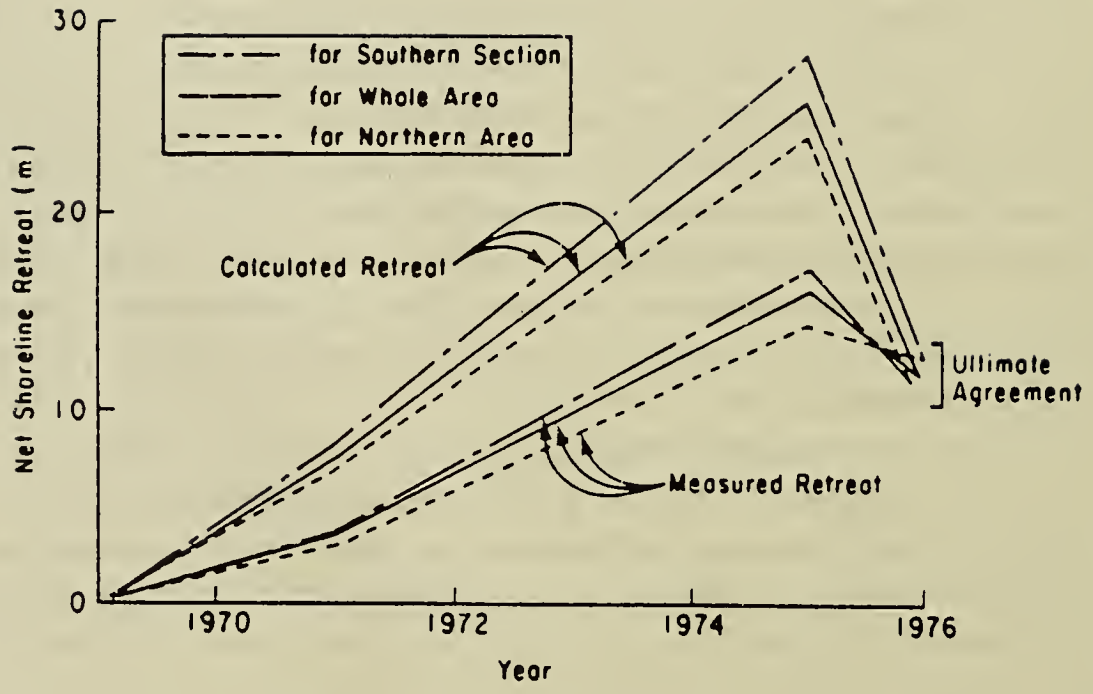


Fig. 7.4. Comparison of Predicted and Measured Shoreline Changes Due to Water Level Increases, Eastern Shore of Lake Michigan (after Hands, 1983). Reprinted with permission from Handbook of Coastal Processes and Erosion. Copyright CRC Press, Inc., Boca Raton, Florida.

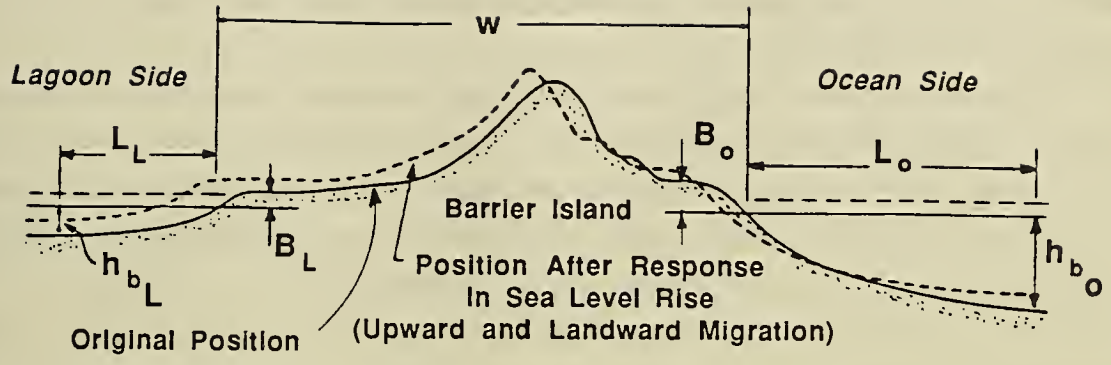


Fig. 7.5. Generalized Shoreline Response Model Due to Sea Level Rise. Applicable for a Barrier Island System which Maintains its Form Relative to the Adjacent Ocean and Lagoon (after Dean and Maurmeyer, 1983). Reprinted with permission from Handbook of Coastal Processes and Erosion. Copyright CRC Press, Inc., Boca Raton, Florida.

width), and 2) the portion of the profile now being "mined" to yield compatible sediment is the difference between ocean and bay depths, $h_{b_o} - h_{b_L}$ (i.e., smaller). This equation simplifies to the Bruun Rule if only the ocean side of the barrier system is active. Finally it is noted that as the bay depth h_{b_L} approaches the active ocean depth, h_{b_o}, Eq. 7.4 predicts an infinite retreat rate. This may explain in part the phenomenon of "overstepping" in which barrier islands, rather than migrating landward retaining their identity in the process, are overwashed and left in place as a linear shoal (see, for example, Sanders and Kumar, 1975).

It is noted that Eqs. 7.3 and 7.4 both consider the portion of the profile being "mined" for sand as containing 100% compatible material. If a portion of the profile contains peat or fine fraction that will not remain in the active system, a rather straightforward modification of the equations is required.

Kriebel and Dean (1985) have described a dynamic cross-shore transport model in which the input includes the time-varying water level and wave height. In addition to predicting long-term responses, this model accounts for profile response to very short-term events such as hurricanes. Thus an equilibrium profile is not assumed and, in addition to a sand budget "volumetric equation," a "dynamic equation," is required which was hypothesized as

$$Q_s = K(D - D_*) \tag{7.5}$$

in which Q_s represents the offshore sediment transport per unit length of beach, K is a universal constant ($K = 2.2 \times 10^{-6}$ m^4/N in the metric system) and D and D_* represent the actual and equilibrium wave energy dissipation per unit water volume. Eq. 7.5 is suggested following the determination by Bruun (1954) and later by Dean (1977) that most equilibrium beach profiles are of the form

$$h = Ax^{2/3} \tag{7.6}$$

in which A is a dimensional profile constant depending primarily on sediment size but secondarily on wave climate. Dean (1977) found that Eq. 7.6 is consistent with uniform wave energy dissipation per unit volume. The quantities A and D_* are related by

$$A = \left[\frac{24}{5} \frac{D_*}{\rho g \sqrt{g} \kappa^2}\right]^{2/3} \tag{7.7}$$

in which ρ is the mass density of water, g is the gravitional constant and κ is the ratio of spilling breaking wave height to water depth ($\kappa \approx 0.8$).

All models of beach profile response described earlier require the identification of a limiting depth of motion h_* in Eq. 7.3 and h_{b_o} and h_{b_L} in Eq. 7.4. Hallermeier (1981) has proposed an approximate method for predicting this depth, h_*, based on average annual significant deep water wave height, $\bar{H}_s$, and period $\bar{T}_s$ and sediment size D,

$$h_* = (\bar{H}_s - 0.3\sigma)\ \bar{T}_s\ (g/5000D)^{0.5} \tag{7.8}$$

in which σ is the standard deviation of the significant wave height.

The models presented heretofore invoke the concept of a limiting depth of motion, a depth seaward of which conditions are static or at least there is no substantial exchange of sediment with the more active shoreface. This assumption seems innocent and quite natural, yet the consequences are very substantial. If no interchange with the shelf profile occurs, erosion is the only possible shoreline response to sea level rise (i.e., there can be no shoreward transport contributions from the continental shelf). There is evidence that shoreward sediment transport is a major contributor to shoreline stability in many areas. The erosion along the south shore of Long Island and at Montauk Point is clearly too small to provide the

well-documented westward net transport at Fire Island (Dean, 1986; Williams and Meisburger, 1987).

Dean (1987) has suggested that during the more rapid rate of sea level rise up to 6,000 year BP, the shoreward shelf transport was not sufficient to maintain a stable shoreline. However, with the relative sudden sea level rise reduction by an order of magnitude, the same rates of shoreward sediment transport generally led to reduced erosion rates and in some cases to stable or accreting shorelines; Fig. 7.6 illustrates the concept. The equilibrium mechanics associated with this concept are much different than those employed by Bruun. Recognized are the natural variability of waves and sediment sizes with sorting resulting in coarser sediment close to shore. It is hypothesized that a particle of a given size is in equilibrium when it is in a certain water depth at a particular distance from shore. With sea level rise and assuming that the wave climate remains the same, the sediment particle would tend to move landward rather than seaward as required by the Bruun Rule. Fig. 7.7 illustrates this mechanism of sedimentary equilibrium.

With greater and greater sea level rise, the general situation will shift toward erosion. Of primary importance is that to predict the response to sea level change, each shoreline segment must be considered on a case-by-case basis with due consideration of the sediment budget. The components of the sediment budget are difficult to quantify. The best basis for developing an appropriate response model for a shoreline segment is an analysis of past response, including a focus on possible anthropogenic effects.

In discussing shoreline response models to sea level change and their development and calibration, it is important to recognize and respect the amount of "noise" in the system including that introduced anthropogenically. Coastal structures and sand management practices at navigational channel entrances are undoubtedly the main contributors to shoreline perturbation by humans. The special attention to documentation following storm activity should also be noted. Along the east coast of Florida, in excess of 38 million cubic meters of beach compatible material has been dredged from channel entrances and disposed at sea. Based on the Bruun Rule, this amount is enough to offset 70 years of shoreline retreat

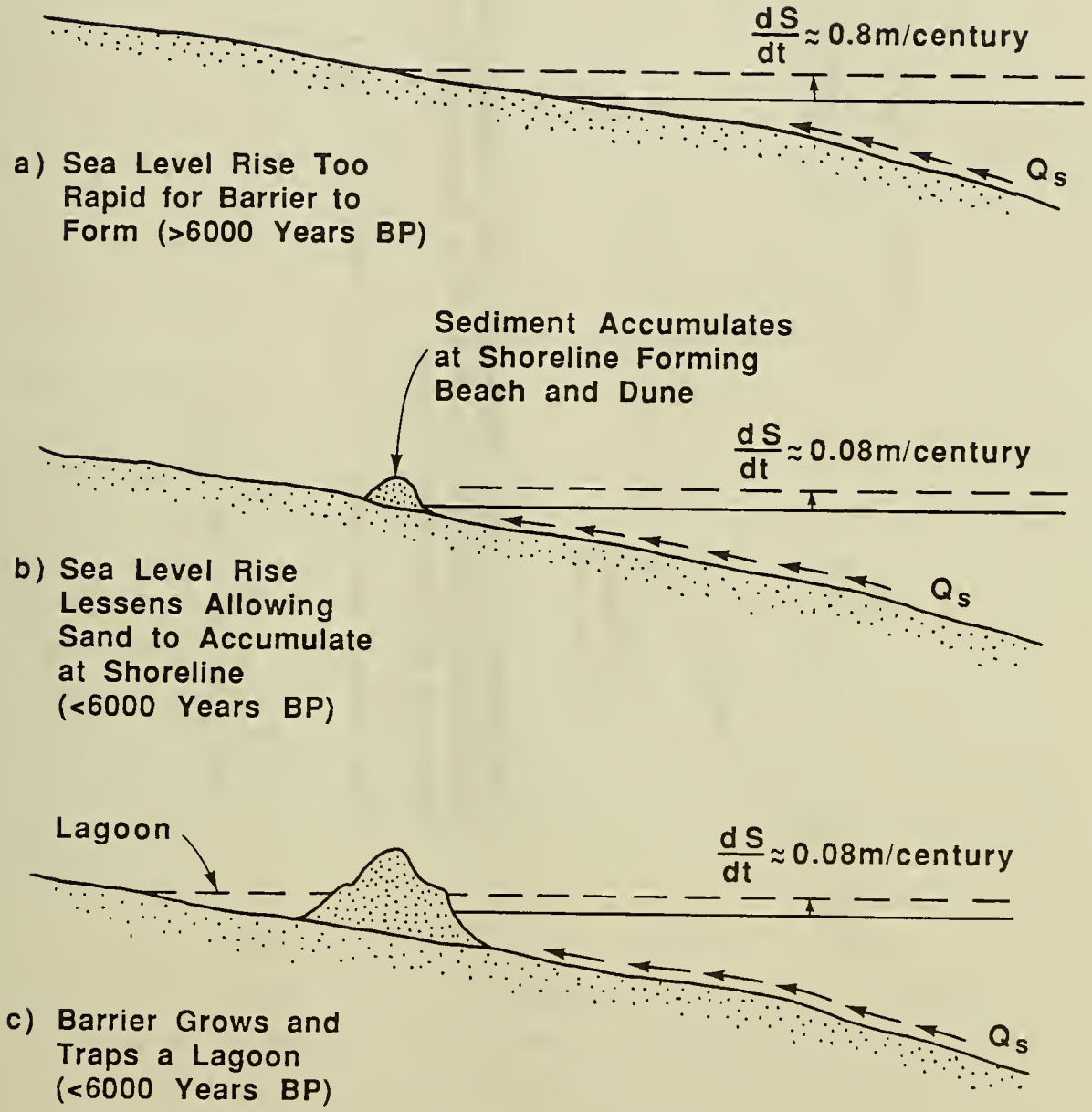
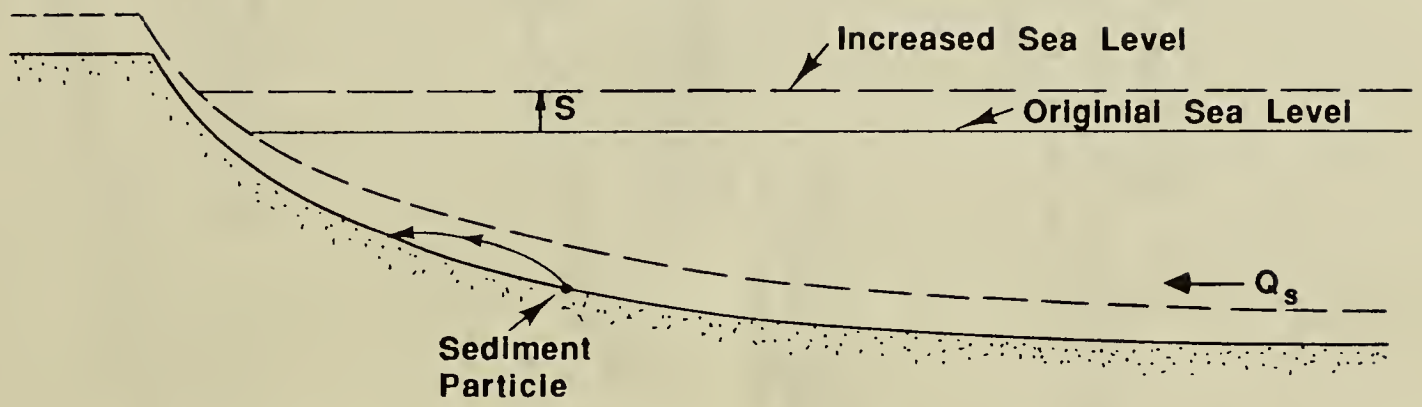


Fig. 7.6. The Role of Shoreward Sediment Transport, Q_S, Across the Shelf and Rate of Sea Level Rise in Causing Barrier Island Formation (from Dean, 1987, courtesy of Shore and Beach).



"Subjected to a Given Statistical Wave Climate, A Sediment Particle of a Particular Diameter is in Statistical Equilibrium When in a Given Water Depth"

Thus When Sea Level Increases, Particle Moves Landward

Fig. 7.7. Possible Mechanism of Sedimentary Equilibrium (from Dean, 1987, courtesy of Shore and Beach).

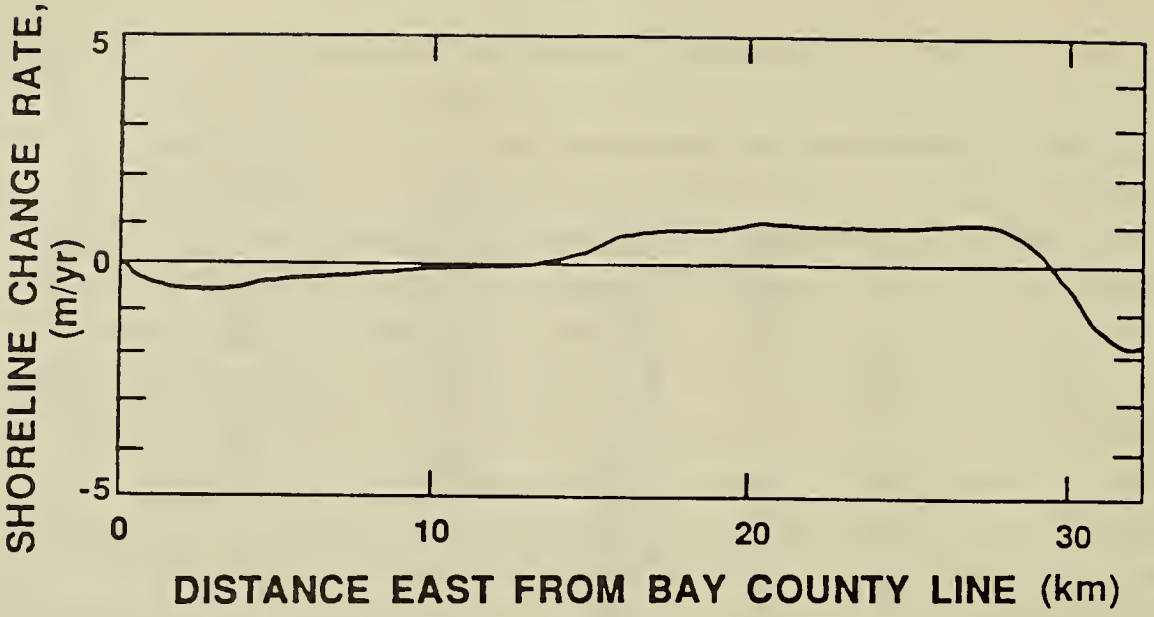
using a eustatic sea level rise of 1.2 mm/year and a retreat/rise multiplier of 100. Data provided by the Jacksonville District of the U.S. Army Corps of Engineers for the period 1980 to 1985 indicate that approximately 50% of the east Florida coast material dredged was still being disposed at sea during this period. This amount (38,000 m^3), again using the Bruun Rule, is sufficient to more than offset their retreat due to the eustatic sea level rise rate employed in the preceding example.

The role of inlets in Florida has been well documented in two cases. The entrance to St. Andrews Bay was cut in 1934 on a previously stable beach. Over the next 50 years, the beach receded at a maximum rate in excess of 2 m/yr where accretion of 1 m/yr had occurred prior to cutting the inlet, Fig. 7.8. The second example illustrates both the adverse effect of cutting the entrance to Port Canaveral in 1951 and the beneficial effects of a beach restoration project carried out in 1974. Again as shown in Fig. 7.9, a beach that had been stable previously underwent dramatic erosion immediately downdrift (south) of the inlet.

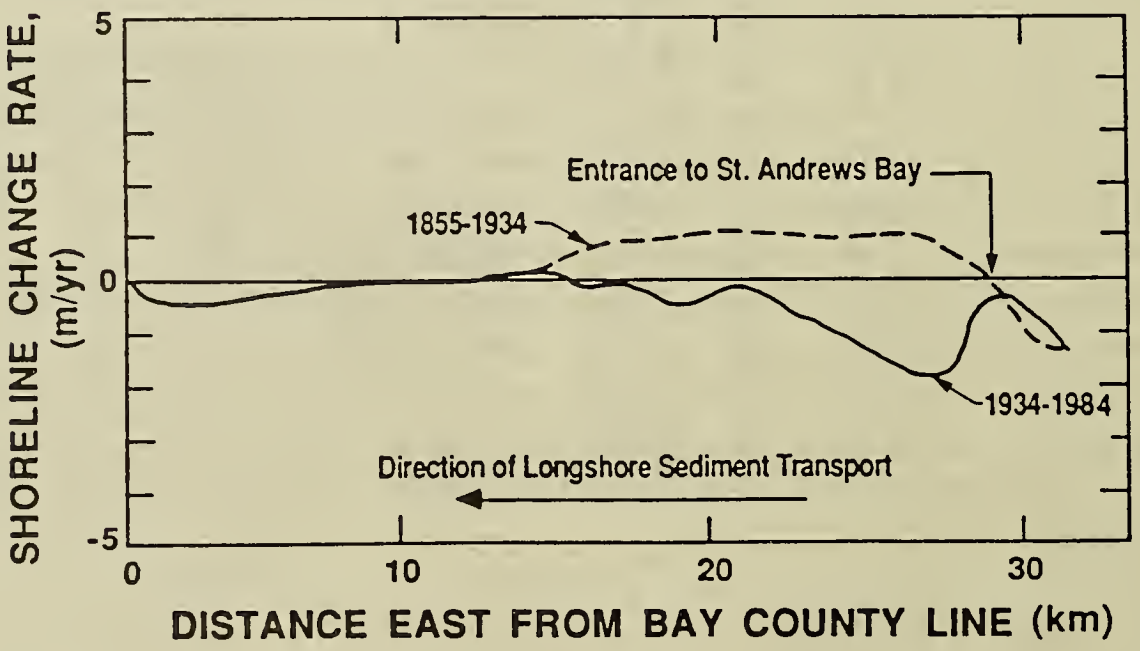
Weggel (1986) has examined the economics of beach nourishment under the scenario of a rising sea level. Methods were presented for computing the present worth costs of perpetual renourishment for sea level rising at a uniform rate and projects of limited life (e.g., 50 and 100 years) for increasing sea level rise rates as predicted by Hoffman (1984). The tradeoff between renourishment (repeated costs) and stabilizing structures (initial cost only) was examined and, based on the reduced required frequency of renourishment due to the structures, the justified cost of structures is presented. It is concluded that perpetual beach nourishment is not economically justified under the sea level rise rates predicted by Hoffman.

7.3 PHYSICAL PRINCIPLES

There are two general types of considerations that can be applied to beach profile response. Kinematic considerations relate to sand budget components regardless of the causes of the transports and associated forces. Dynamic considerations relate to the forcing mechanisms. Each of these will be discussed briefly in the paragraphs below.

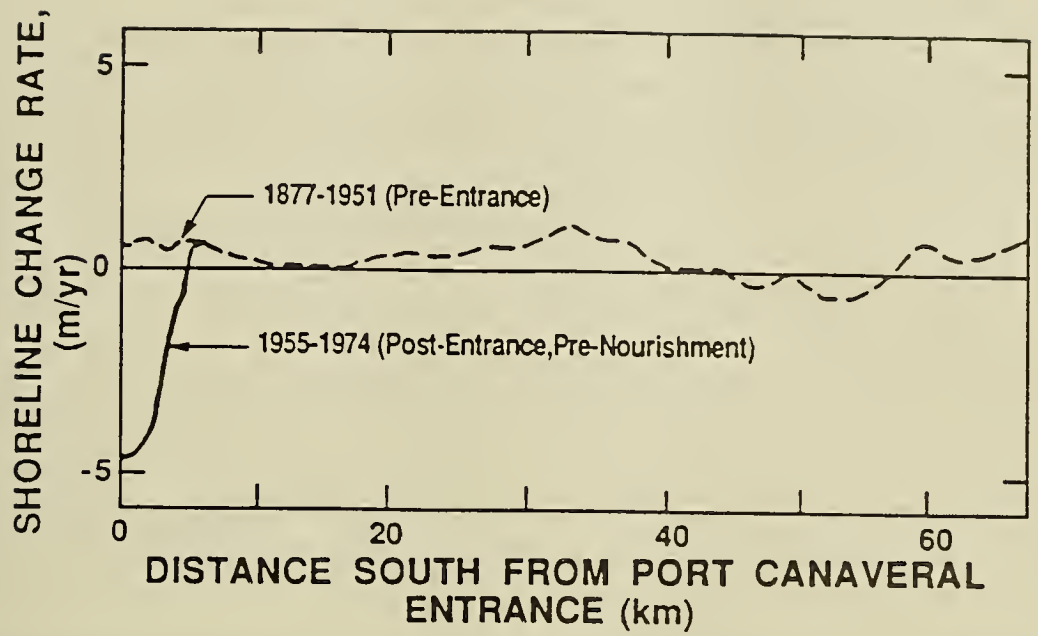


a) Shoreline Change Rates Prior to Cutting Entrance to St. Andrews Bay, 1855-1934 (79 Years).

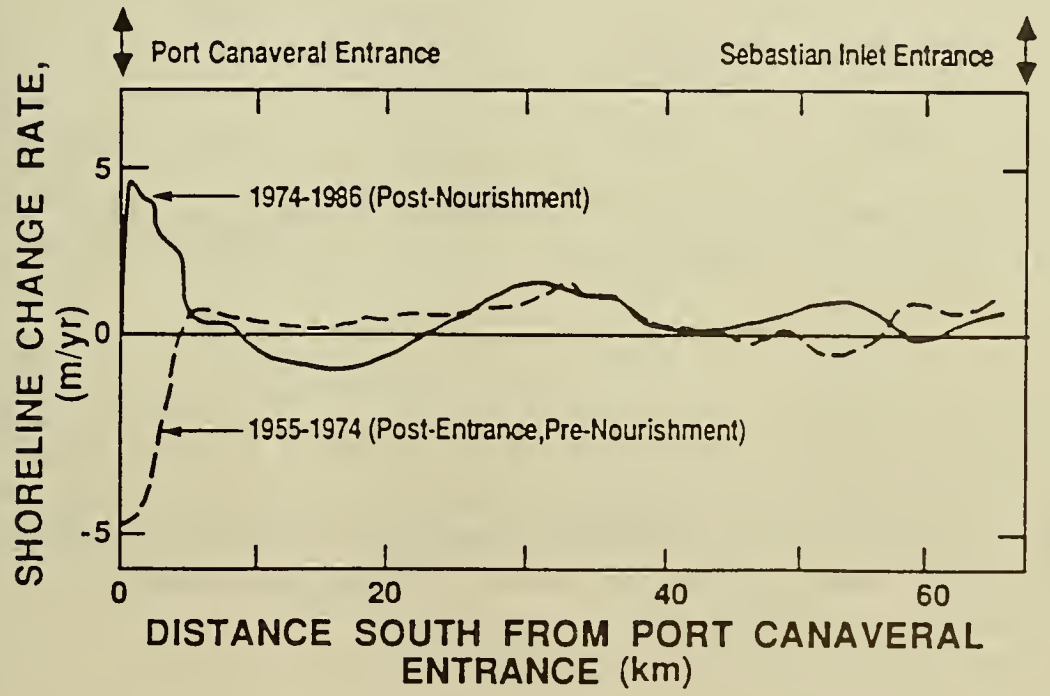


b) Comparison of Shoreline Change Rates Prior to Cutting Entrance to St. Andrews Bay, 1855-1934 (79 Years) and Subsequent to Cutting Entrance, 1934-1984 (50 Years).

Fig. 7.8. Effect of Cutting Entrance to St. Andrews Bay in 1934 on Downdrift Shoreline (from Dean, 1987, courtesy of Shore and Beach).



a) Effects of Channel Entrance on Downdrift Beach Stability, Compared to Pre-Entrance Condition.



b) Shoreline Changes Following 1974 Nourishment Project.

Fig. 7.9. Effects of Establishment of Cape Canaveral Entrance and Subsequent Nourishment Project on Downdrift Beaches (from Dean, 1987, courtesy of Shore and Beach).

7.3.1 Kinematic (Sediment Budget) Considerations

The change in absolute elevation, z, of the sediment-water interface at a horizontal point (x,y) can be expressed in terms of the sediment transport components, q_x and q_y and any source terms,

$$\frac{\partial z}{\partial t} = - \left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y}\right) + s \tag{7.9}$$

For reference purposes, the x and y coordinates are selected in the offshore and alongshore directions, respectively such that q_y represents the local longshore component of sediment transport. The formal origin of the Bruun Rule can be demonstrated by assuming that there are no sources ($s = 0$), that there are no gradients in longshore sediment transport $\partial q_y/\partial y = 0$, and integrating the resulting equation from x_1 to x_2,

$$\int_{x_1}^{x_2} \frac{\partial z}{\partial t}\, dx = - \int_{x_1}^{x_2} \frac{\partial q_x}{\partial x}\, dx = q_{x_1} - q_{x_2} \tag{7.10}$$

and if q_{x_1} and q_{x_2} are sufficiently landward and seaward that no motion occurs, $q_{x_1} = q_{x_2} = 0$, resulting in

$$\int_{x_1}^{x_2} \frac{\partial z}{\partial t}\, dx = 0 \tag{7.11}$$

We now decompose z into two components, one due to the uniform vertical profile displacement, z_1, and the second to the uniform horizontal displacement $z_2(x)$, that is,

$$z = z_1 + z_2(x) \tag{7.12}$$

and Eq. 7.11 becomes

$$\int_{x_1}^{x_2} \frac{dz}{dt}\, dx = \int_{x_1}^{x_2} \frac{dz_1}{dt}\, dx + \int_{x_1}^{x_2} \frac{dz_2}{\partial t}\, dx = 0 \tag{7.13}$$

Since $z_1 \neq f(x)$ and in fact $\partial z/\partial t = \partial S/\partial t$ over the active profile length, L, and by the chain rule,

$$\frac{\partial z_2}{\partial t} = \frac{\partial z_2}{\partial x}\,\frac{\partial x}{\partial t} \tag{7.14}$$

and $\partial x/\partial t$ is uniform over depth and equal to $-\, dR/dt$, Eq. 7.13 becomes

$$\frac{\partial S}{\partial t}\, L = \frac{\partial R}{\partial t} \int_{z_1(x_1)}^{z(x_2)} \frac{\partial z}{\partial x}\, dx \tag{7.15}$$

and the active depth is h_*+B, we finally obtain

$$\frac{\partial R}{\partial t} = \frac{\partial S}{\partial t}\,\frac{L}{(h_*+B)} \tag{7.16}$$

which is the Bruun Rule.

We now return to a somewhat more conceptual version of the cross-shore sediment budget equation, expanding the source term. Considering the case of shoreline stability for illustration purposes

$$\frac{dz}{dt} = \frac{dS}{dt} = -\frac{\partial q_x}{\partial x} + SS + B \qquad (7.17)$$

in which the first term on the right hand side is the local convergence of cross-shore bottom sediment transport, the second term, SS, represents deposition of suspended sediment and the last term is the local biogenic production. The required magnitudes of each of these in order to compensate for a eustatic sea level rise of 1.2 mm/yr alone is presented below.

Convergence of Cross-shore Sediment Transport - Considering a shelf width of 40 km

$$q_{x_2} = 40{,}000 \text{ m}(0.0012 \text{ m/yr}) = 48 \text{ m}^2/\text{yr} \qquad (7.17a)$$

Suspended Sediment Deposition

$$SS = 0.0012 \text{ m/yr} \qquad (7.17b)$$

Biogenic Production

$$b = 0.0012 \text{ m}^3/\text{m}^2/\text{yr} \qquad (7.17c)$$

At present, knowledge is extremely poor concerning the rates of the three components addressed above. However, until improved quantification is available, all should be regarded as potentially substantial contributors to the shoreface sediment budget.

7.3.2 Dynamical Considerations

It is clear that there are net shoreward directed hydrodynamic forces acting on a sediment particle; otherwise an upward sloping beach could not be in equilibrium against the forces of gravity. The near-bottom flow field due to a storm system is complex and consists of net forces and

destabilization forces. The dominant destabilization forces are due to the combined effects of gravity and due either to offshore directed flows in water depths greater than breaking or due to wave breaking in the shallower depths. Fig. 7.10 presents the general situation.

The constructive effects include the shoreward predominance in bottom shear stress due to nonlinear waves. The oscillatory bottom water particle velocity, u, associated with a nonlinear wave can be expressed as

$$u = a_1 \cos\alpha + a_2 \cos 2\alpha + \ldots \tag{7.18}$$

in which α is the phase angle and the a_n are velocity amplitude coefficients. Even though the <u>time</u> averaged bottom velocity is zero ($\bar{u} \equiv 0$), the net onshore shear stress, $\bar{\tau}$, is positive (i.e., shoreward) since

$$\tau = \frac{\rho f}{8} |u|u \tag{7.19}$$

where ρ is the mass density of water and f is the Darcy-Weisbach fraction factor. Fig. 7.11 is based on stream function (nonlinear) wave theory and presents the average nondimensional shoreward shear stress, $\bar{\tau}'$,

$$\bar{\tau}' = \frac{\overline{|u|u}}{(H/T^2)} \tag{7.20}$$

as a function of relative water depth, h/L_o and wave steepness, H/L_o.

7.4 RESEARCH NEEDS

It is clear that the development of an adequate capability to predict shoreline response to future sea level rise rates will require a consideration of cross-shore sediment transport fundamentals and applications, and a quantitative understanding of the transport components. The Bruun Rule, while a good first model, is deficient in not allowing for the onshore transport of sand that is clearly occurring at some locations and undoubtedly occurring at many less evident locations. The three types

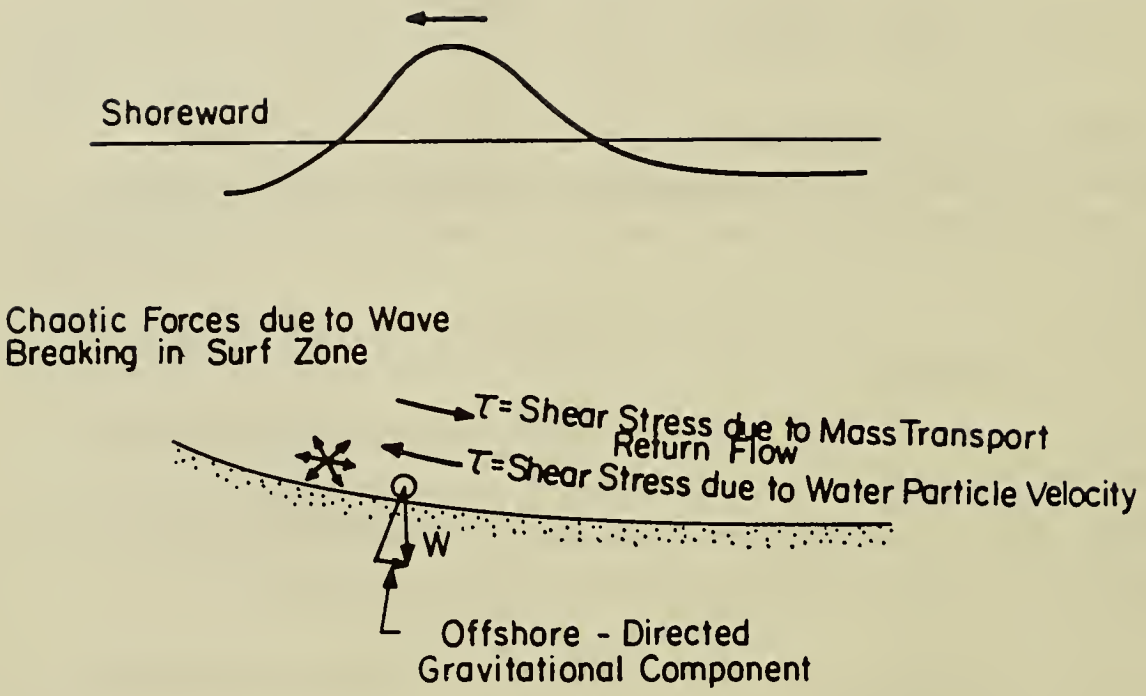


Fig. 7.10. Dominant Forces Acting on a Sediment Particle Resting on the Bottom.

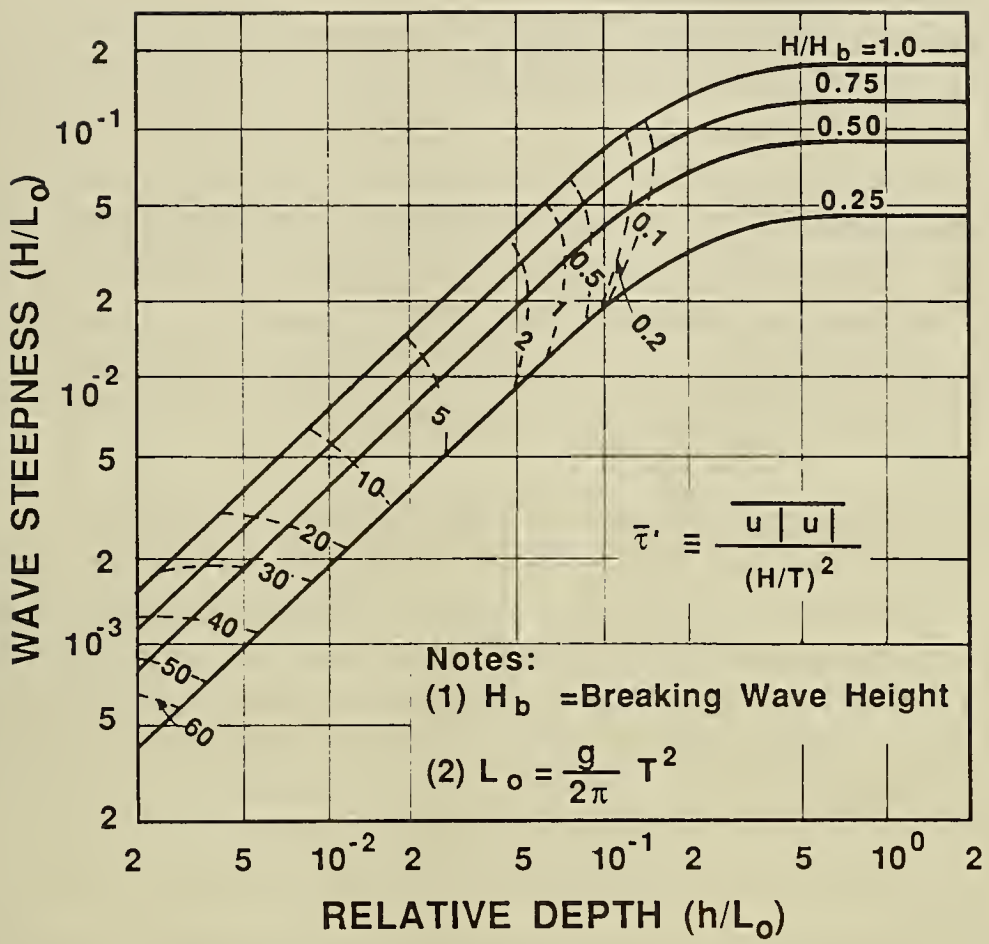


Fig. 7.11. Isolines of Non-dimensional Average Bottom Shear Stress $\overline{\tau'}$ vs Relative Depth h/L_o, and Wave Steepness, H/L_o (from Dean, 1987, courtesy of Shore and Beach).

of research needs identified fall in the categories of analysis of existing data, new data, and new technology.

7.4.1 Analysis of Existing Data

Isolation of Anthropogenic Effects - The substantial effects that navigational structures and sand management practices at entrances can have on shoreline stability have been noted and illustrated by the examples in Figs. 7.8 and 7.9. Similar effects are known to occur and be substantial at many other locations, i.e., Folly Beach, SC, Tybee Island, GA, Santa Barbara, CA, and Assateague Island, MD. In addition to the effects at entrances, the effects of groins, seawalls, etc. should be considered. A straightforward methodology could be applied; however, it is believed that development of new, more effective methodology would be worthwhile.

Regional Correlation of Shoreline and Sea Level Change Rates - Previous estimates of long-term shoreline change have been developed (U.S. Army Corps of Engineers, 1973; Dolan et al., 1983). These estimates are available on a state wide basis and regionally (such as the entire Atlantic coast). It would be a useful and instructive first broad-brush effort to correlate these estimates with local estimates of sea level rise over the past 50 years or so.

7.4.2 New Data

Quantification of Cross-shore Sediment Budget Components - Referring to Eq. 7.17, the focus of this research element would be to quantify the three terms on the right hand side. The complete methodology has not been developed as yet, but would probably consist of long-term observations of offshore stakes to determine total vertical change, studies of biogenic production and attrition and deposition rate by suspended sediment traps. It would be useful to conduct this element in conjunction with the experimental element of "Evidence from the Continental Shelf" to be described below.

Shoreline Monument System - The state of Florida maintains a monumented baseline around 1,030 km of sandy shoreline. Since the early to

mid 1970s, comprehensive surveys have been conducted on approximately a decadal basis and post-storm studies carried out when appropriate. This type of system provides the only basis for obtaining quality information of shoreline change. It would be very worthwhile, in anticipation of the rising concern over shoreline response to sea level rise, to encourage other states to install, monitor and maintain a monumented system similar to that of Florida.

Evidence from the Continental Shelf - The seafloor of the continental shelf contains information relating to past shoreline response to sea level rise and potential future response. Specifically, Swift (1975) has shown that along much of the Mid-Atlantic Region, there is a "lagoonal carpet" of muds that could not contribute significantly to the sediment budget of the active shoreface. Additionally, the shape of the offshore profiles, along with the availability of sand-sized material contains information (although as yet not completely understood) whether the offshore profile will serve as a source or sink of sand.

In addition to the above, it would be worthwhile to conduct measurements of long-term sediment movement on the continental shelf. These measurements would be conducted along a representative profile; they would document the forcing function (waves, currents, tides, stresses, etc.) and sediment transport (response function). The sediment transport would best be documented through passive means, such as the use of sediment tracing techniques.

7.4.3 New Technology

Laser Profiling - A laser profiling system has been developed in which the airborne laser oscillates in a conical pattern thereby sweeping out a swath of one-half the airplane height, with the ground level pattern being nearly circular overlapping trajectories. The laser return establishes the dry beach elevation, the water surface and the below water profile to approximately two "secchi" disk depths. The potential of this technology to systematically and periodically conduct regional and tidally controlled surveys of shoreline change is extremely encouraging. It is recommended that a pilot program be conducted in Florida in the coming year.

8. SALTWATER INTRUSION

8.1 INTRODUCTION

Many coastal cities rely on local groundwater to meet domestic and industrial needs. With the increasing demand due to greater population and industrial concentrations along the coastline coupled with sea level rise, the potential for saltwater contamination of the aquifer will increase. As in other cases, there are two general approaches to responding to this problem: 1) abandonment of use of the resource, or 2) adoption of management and prevention measures to reduce salinity intrusion. Considerable experience has been gained in coping with saltwater intrusion not principally due to sea level rise, but due to excessive use of the groundwater resource. Yet sea level rise and excessive groundwater usage both decrease the seaward directed piezometric gradient and are, in some respects, comparable. This chapter reviews the various types of saltwater problems that can occur due to sea level rise and the capability to predict and respond to such intrusion.

8.2 LITERATURE REVIEW

The subject of saltwater intrusion into coastal aquifers has been of interest for several centuries. Recently the interest has increased due to intense use of coastal groundwater resources. Methods have been developed to predict the effects of differing usage and management procedures.

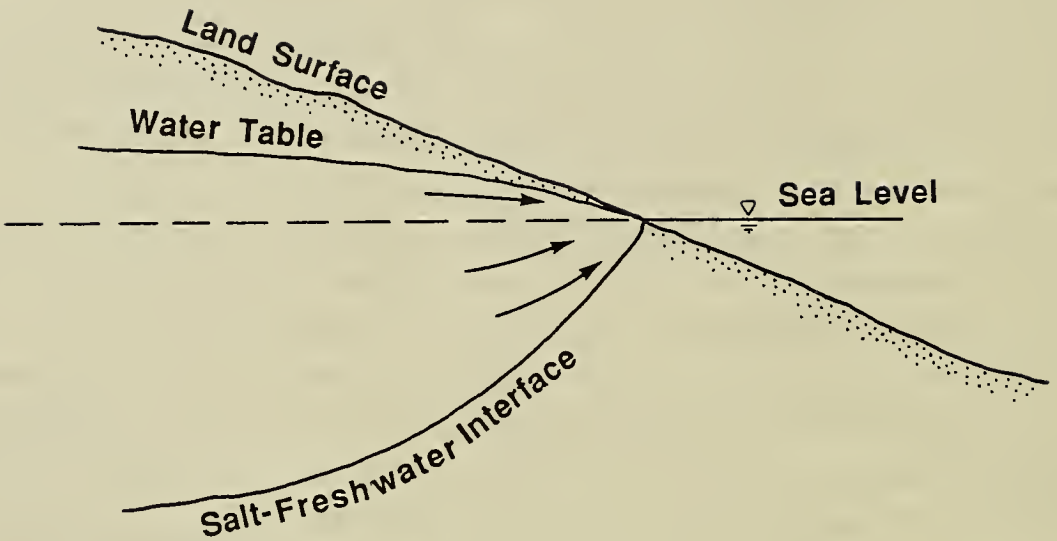
Todd (1980) has presented a review of the theory and management practices related to saltwater intrusion in coastal aquifers. In addition to the excessive pumping and sea level rise, saltwater intrusion can result from surface drainage canals which both lower the freshwater head, and if salt water flows into the canals, surface contamination. In recognition of this problem, the State of Florida has enacted legislation to establish a saltwater barrier line in areas prone to saltwater intrusion through canals (Hughes, 1979). There are a number of approaches for controlling saltwater intrusion, as summarized in Table 8.1. Several of these techniques will be illustrated later by examples.

Table 8.1. Methods for Controlling Saline Water Intrusion (reprinted with permission from D. K. Todd, Groundwater Hydrology, copyright 1980, John Wiley & Sons, Inc.)

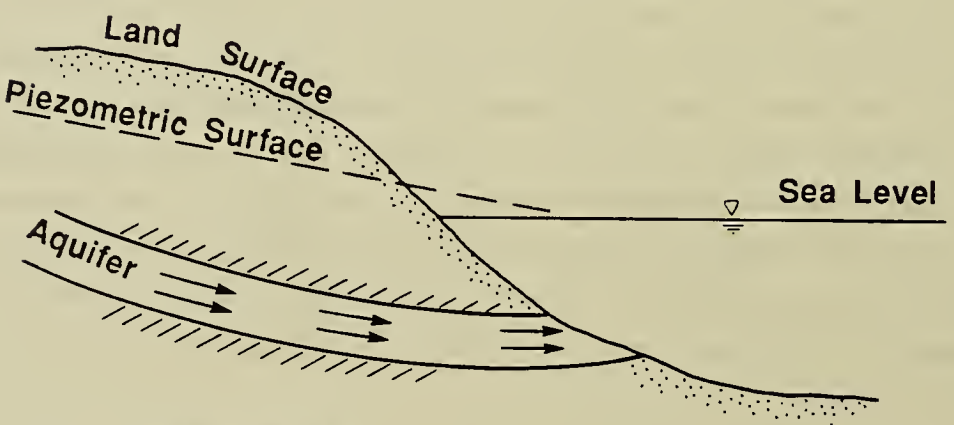
Source or Cause of Intrusion	Control Methods
Seawater in coastal aquifer	Modification of pumping pattern Artificial recharge Extraction barrier Injection barrier Subsurface barrier
Upconing	Modification of pumping pattern Saline scavenger wells
Oil field brine	Elimination of surface disposal Injection wells Plugging of abandoned wells
Defective well casings	Plugging of faulty wells
Surface infiltration	Elimination of source
Saline water zones in freshwater aquifers	Relocation and redesign of wells

Methods of predicting saltwater intrusion are based on Darcy's law (e.g., Todd, 1980) and include analytical (Henry, 1959; Henry, 1964; van der Meer, 1978; Kozeny, 1953; Strack, 1976; Hunt, 1983) and numerical (McDonald and Harbaugh, 1984; Lennon et al., 1987; Atkinson et al., 1986; Shrivastava, 1978) models. The analytical models are useful for obtaining insight into the interrelationship of various parameters, whereas the numerical models are valuable for predicting detailed response for a particular situation including the benefits of various management strategies. The capabilities of numerical modeling appear to be limited only by knowledge of the transmissive and porosity characteristics of the medium, and the availability of the forcing function, and boundary conditions such as inflow at the boundaries. Most of the efforts conducted to date have not been directed toward evaluating the effects of sea level changes; however, the capability appears to exist. The status of saltwater intrusion in aquifers in all 48 coastal and inland states is presented by Atkinson et al. (1986) as a figure for each state.

In discussing the effects of sea level rise on saltwater intrusion, the distinction between unconfined and confined aquifers is important, see Fig. 8.1. Unconfined aquifers that discharge at or near the shoreline are much more vulnerable to saltwater intrusion than are confined aquifers. The idealized landward displacement of the salt-freshwater interface can be predicted based on the sea level rise, the form of the interface at the level of interest and the slope of the land. For confined aquifers, the effect of sea level rise is to cause a feedback which tends to offset the rise. The increased head due to the rise at the point of discharge causes a decrease in the piezometric gradient, a reduction in the discharge rate and a resultant transient that, for the same recharge rate, causes an increase in the inland head. The net effect is to reestablish the same discharge rate and same relative (to sea level) piezometric head as before sea level rise.



a) Unconfined Aquifer



b) Confined Aquifer

Fig. 8.1. Example of Unconfined and Confined Aquifers.

8.3 PHYSICAL PRINCIPLES AND SOLUTIONS TO IDEALIZED PROBLEMS

8.3.1 General

A simple hydrostatic analysis of the balance between a freshwater aquifier meeting a saline water body yields the Ghyben-Herzberg principle

$$z = \frac{\rho_f}{\rho_s - \rho_f} h \tag{8.1}$$

as illustrated in Fig. 8.2. The above relation applies for a distinct fresh-saltwater interface. However, field measurements have proven definitively that the transition between salt and fresh water is quite gradual with mixing occurring over this transition zone, see Fig. 8.3. The principal cause of the gradual transition, which can be interpreted as dispersion, is the movement of this interface back and forth due to the relatively short period astronomical tide components and also the longer term oscillations due to seasonal variations in replenishment by rainfall and the still more infrequent droughts. During the saltwater advancement and retreat, some salt water is left in the interstices. This is the so-called convective mode. The mixing with the retained fresh water occurs by more slowly occurring molecular diffusion. Experiments have been conducted which demonstrate that the dispersion due to ocean tides can be expressed as

$$D = 4MA/t_o \tag{8.2}$$

where M is a parameter with dimensions of length and A and t_o are the horizontal amplitude of tidal motion and period, respectively. Laboratory studies have been shown that the parameter, M, increases with the uniformity of the stratum and can range from 0.063 cm to possibly as high as 2.8 cm for very uniform sand. Methods employing numerical models combined with Taylor's hypothesis relating the dispersion coefficient tensor, the pore water velocity fluctuations and the Lagrangian time scale tensor have been used to predict anisotropic dispersion coefficients (e.g., Fattah and Hoopes, 1985; Chin, 1986). Based on realistic values of tidal and ground permeability characteristics, it can be shown that the

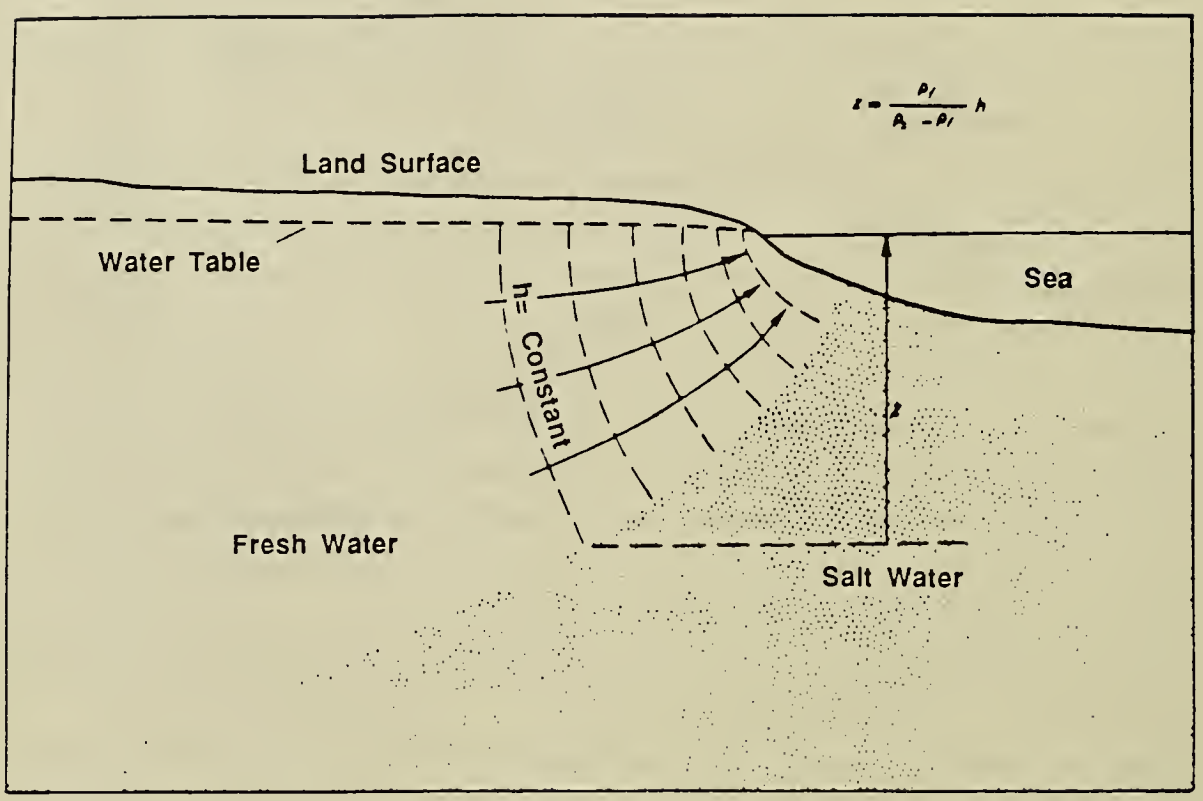


Fig. 8.2. Balance between Fresh Water and Salt Water in a Coastal Aquifer in which the Salt Water is Static (after Cooper, 1964).

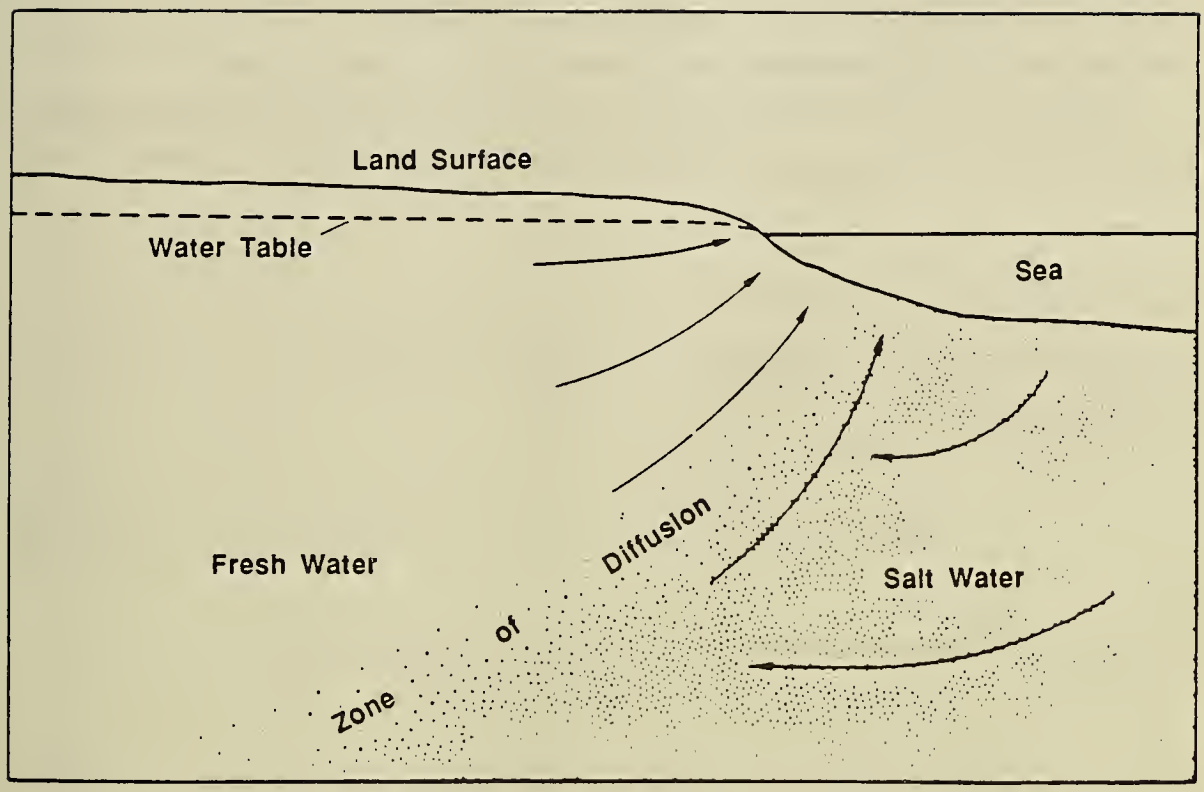


Fig. 8.3. Circulation of Salt Water from the Sea to the Zone of Diffusion and Back to the Sea (after Cooper, 1964).

dispersion coefficient decreases from about 100 cm^2/day to 10 cm^2/day at distances of 90 and 275 m from the shoreline, respectively. At a distance of 425 m, the tidally-induced dispersion coefficient would be about the same as molecular diffusion for sodium chloride.

The well-known Dupuit approximation is based on the assumption that the hydraulic head is constant along any vertical line throughout the water body and is therefore given by the elevation of the free surface. Strictly speaking, although the Dupuit approximation allows only horizontal flow, it provides meaningful solutions to a number of practical problems.

The entire field of analytical and numerical modeling of groundwater flows is based on Darcy's law which can be expressed as

$$\vec{u} = - \vec{\nabla}\phi \tag{8.3}$$

in which $\vec{u}$ represents the three-dimensional velocity vector, ϕ is the velocity potential

$$\phi = K \left(\frac{p}{\rho g} - y\right) \tag{8.4}$$

$\vec{\nabla}$ is the three-dimensional vector differential operator, and

$$\vec{\nabla}() \equiv \vec{i} \frac{\partial}{\partial x} () + \vec{j} \frac{\partial}{\partial y} () = \vec{k} \frac{\partial}{\partial z} () \tag{8.5}$$

ρ is the pressure, y is the depth below a datum, K is the hydraulic transmissivity of the medium, and $\vec{i}$, $\vec{j}$, $\vec{k}$ are the unit vectors in the x, y, and z directions, respectively. The continuity equation for an incompressible fluid is

$$\vec{\nabla} \cdot \vec{u} = 0 \tag{8.6}$$

which, when combined with Eq. 8.3, yields the Laplace equation in ϕ

$$\nabla^2 \phi = 0 \tag{8.7}$$

In two-dimensional flow, the Laplace equation is also satisfied in terms of the stream function, ψ, so that a field of ϕ and ψ can be constructed to represent the usual flow net with lines of constant ϕ representing lines of equal head and lines of constant ψ being streamlines and everywhere perpendicular to lines of ϕ.

With the governing equations noted above, many solutions to practical problems can be obtained through conformal mapping procedures. For particular geometries, possibly time-varying boundary conditions, and complicated flow boundaries, numerical solutions are applied. The U.S. Geological Survey developed and made available (McDonald and Harbaugh, 1984) a numerical model to simulate groundwater flow in three dimensions.

8.3.2 Discharge through an Unconfined Aquifer

Kozeny (1953) has presented a solution which has been modifed for flow conditions from an unconfined aquifer to a shoreline. The solution is developed in terms of complex variables and a distinct interface between salt and fresh water and yields several results of interest. Denoting x and y as the inland and downward coordinates (see Fig. 8.4), respectively, the interface between fresh and sea water is expressed by

$$y = \sqrt{\frac{2Qx}{\gamma K} + \frac{Q^2}{\gamma^2 K^2}} \tag{8.8}$$

in which Q = freshwater flow per unit length of shoreline, and $\gamma = (\rho_s - \rho_f)/\rho_f$. The horizontal width, x_o, through which the fresh water flows to the sea is

$$x_o = -\frac{Q}{2\gamma K} \tag{8.9}$$

The variation of freshwater head, h, with distance from the shoreline is

$$h = \sqrt{\frac{2\gamma Qx}{K}} \tag{8.10}$$

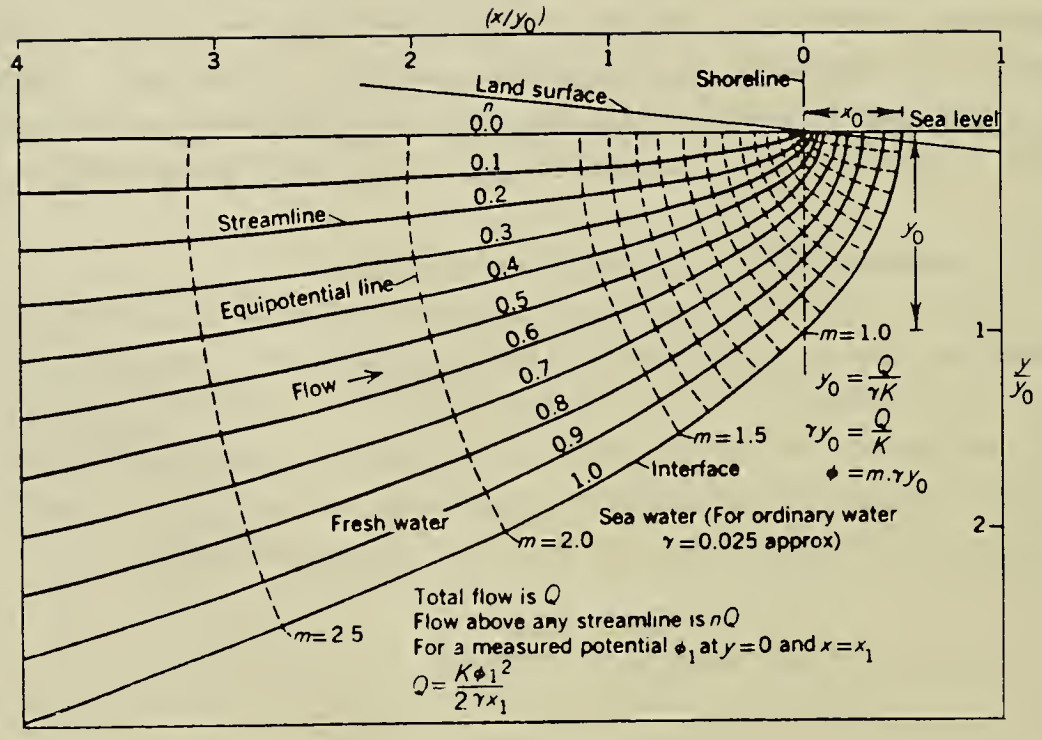


Fig. 8.4. Idealized Characteristics for Unconfined Flow to a Shoreline (after Glover, 1964).

One result of interest is that if the discharge, Q, diminishes due to a drought, groundwater interception by pumping, or other cause, the outflow region, x_o, is reduced so that the aquifer tends to conserve the freshwater region. Also, from Eq. 8.8, it is possible to calculate the increase in elevation of the salt-freshwater interface if Q is decreased. This is relevant to a pumping field with intakes at a particular depth.

As shown in Fig. 8.5, for an unconfined aquifer, the effect of a sea level rise would be to displace the fresh-saltwater interface landward a distance dependent on the slope, s, of the topography. Thus for a sea level increase, S, the interface would be elevated by a distance S and displaced landward a distance, Δx, where approximately

$$\Delta x = \frac{S}{s} \tag{8.11}$$

For a confined aquifer, which flows to the ocean under pressure, saltwater intrusion into the aquifer is not expected to present the same magnitude of problem as for the unconfined aquifer. As shown in Fig. 8.1b, the only requirement is that the piezometric head at the point of aquifer outcropping must be equal to the depth of water at this point. With an increase in sea level, the rate of outflow can decrease, thereby increasing the piezometric head at the point such that little if any intrusion will occur.

8.3.3 Oceanic Islands

Many oceanic islands are composed of relatively permeable limestone with a freshwater layer overlying a saltwater layer. The freshwater layer must be maintained by rainfall or the layer thickness would diminish to zero with increasing time. Referring to Fig. 8.6, the freshwater layer must outcrop at or below the mean sea level.

Considering a circular island of radius, R, with a rainfall recharge rate, W, and employing the Dupuit and Ghyben-Herzberg relations, an approximate freshwater boundary can be determined. The distance, z, to the interface is given by

$$z^2 = \frac{W(R^2 - r^2)}{2K[1 + \frac{\Delta\rho}{\rho}][\frac{\Delta\rho}{\rho}]} \tag{8.12}$$

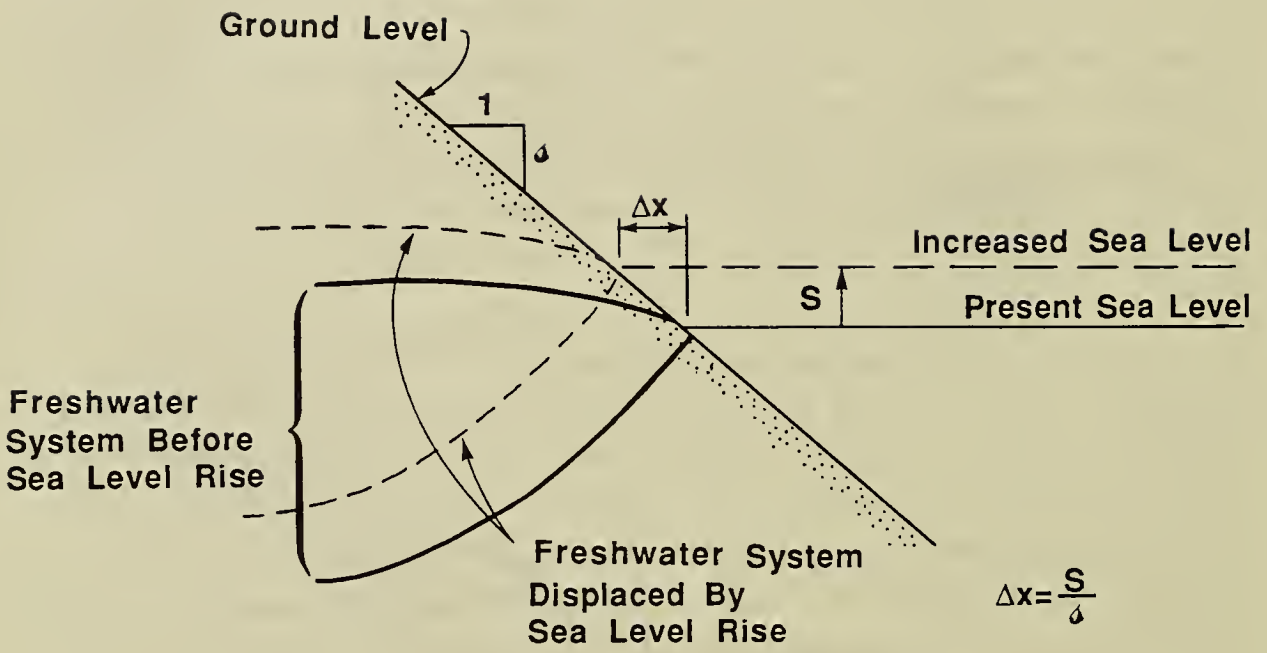


Fig. 8.5. Effect of Sea Level Rise on Equilibrium Groundwater, Highly Exaggerated Vertical Scale.

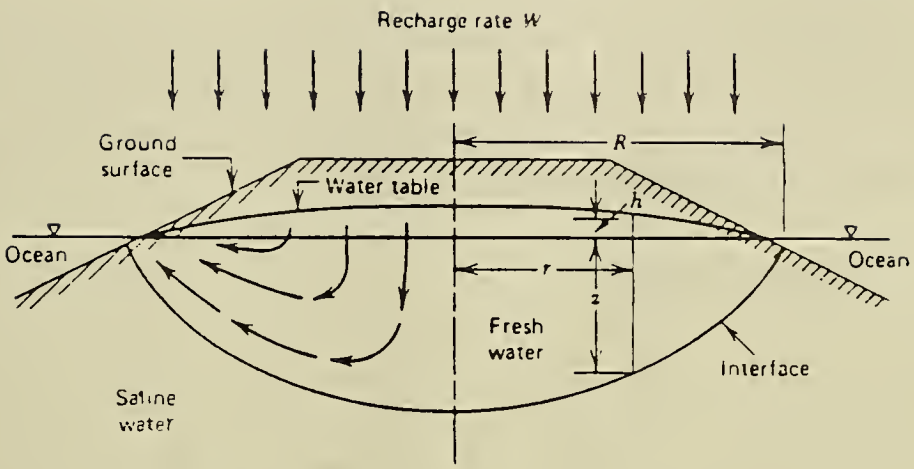


Fig. 8.6. Freshwater Lens Under a Circular Oceanic Island Under Natural Conditions (reprinted with permission from D. K. Todd, Groundwater Hydrology, copyright 1980, John Wiley & Sons, Inc.).

in which K is the hydraulic conductivity and the remaining terms are given in Fig. 8.6. Eq. 8.12 presents the idealized situation for a sharp interface; however as noted before, there will always be a fairly gradual transition from fresh to salt water.

The approach yielding the results presented above for a circular island can be applied also to represent a two-dimensional peninsula.

8.3.4 Upconing

One possible effect of groundwater pumping is that "upconing" of a lower saltwater layer may occur in which the decreased pressure caused by pumping draws the salt water into the intake. The maximum discharge, Q_{max}, by a single well to avoid upconing is given by

$$Q_{max} \leq \pi d^2 K(\Delta\rho/\rho) \tag{8.13}$$

in which d is the distance from the bottom of the well to the static interface.

With increasing sea level, assuming that rainfall recharge and other parameters are unchanged, the freshwater region portrayed in Fig. 8.6 will be elevated by the same amount as the sea level rise; and with the same absolute intake depth, the potential for upconing would increase. However, by elevating the well intake point the same amount as sea level rise, the upconing situation would be unchanged to a first approximation from that with the lower sea level. Increased groundwater pumping in amounts greater than the net renewal rate to the aquifer will elevate the interface. An example is at Honolulu where the depths of production wells have decreased gradually from 450 m to 85 m as the freshwater layer thickness has been decreased by excessive pumping.

8.3.5 Single Extraction Well Near a Coast

An analytical solution has been developed by Strack (1976) to represent the flow field in a confined aquifer in the vicinity of a coastline as affected by the withdrawal from a single well, see Fig. 8.7. For the case in which the unperturbed flow per unit length of coastline is

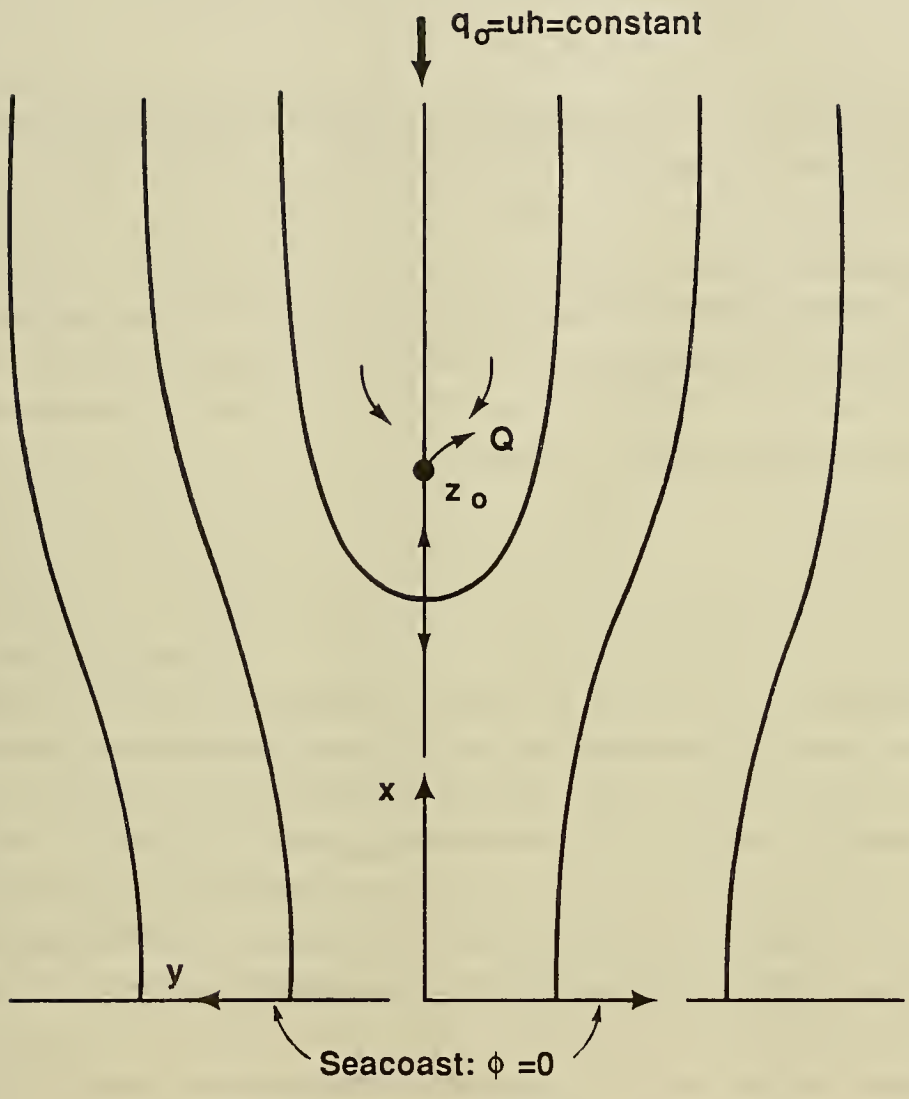


Fig. 8.7. Flow to a Single Well Along a Seacoast.

q and with the well located a distance x_o from the coast, the critical withdrawal, Q_c, is defined by the following two equations

$$\frac{KD^2}{q_o x_o} \gamma(1 + \gamma) = 2 \frac{x}{x_o} - [1 - (\frac{x}{x_o})^2] \ln [\frac{1 + x/x_o}{1 - x/x_o}]] \quad (8.14)$$

$$\frac{Q}{q_o x_o} = \pi [1 - (\frac{x}{x_o})^2] \quad (8.15)$$

In the above equations, D is the vertical distance from mean sea level to the aquifer bottom boundary.

8.3.6 Saltwater Barriers

In areas where solution channels exist that convey large quantities of fresh water to sea, the possibility exists of employing underwater dams. This approach has been tested successfully in the limestone caverns of Port-Miou River near Marsielles, France.

8.4 CASE STUDIES

There are two general approaches for reducing saltwater intrusion into coastal aquifers: 1) modifying pumping practice, and 2) construction of flow barriers. The most simple and direct approach is, where appropriate, to reduce groundwater withdrawal during periods of drought when intrusion would tend to occur. It is clear that with increased sea level, the frequency and/or duration that intrusion would tend to occur will increase. Unfortunately, drought occurrences are precisely the times that demand on groundwater resources tends to be increased. An alternate approach is to adopt a pumping plan to minimize intrusion through hydraulic or structural barriers.

Five case studies will be presented of approaches to cope with salinity intrusion into aquifers.

8.4.1 Long Island, NY

Todd (1980) has described the effects of excessive pumping near the more populated western end of Long Island including a reduction in the water table to 10 m below sea level. The response has been abandonment of

well fields on the western end of the island and an effort to salvage storm water and recharge it into the aquifer through infiltration fields. The present horizontal rate of advance of the saline wedge in southwestern Nassau County varies from 3 m to 60 m per year depending on local pumping rates and recharge by variations in rainfall. Fig. 8.8 presents a cross section through southwestern Nassau County.

8.4.2 Miami, FL

The initial cause of salinity intrusion in this area was a series of surface drainage canals. These canals both lowered the fresh water table as they were designed to do and also allowed salt water to penetrate up these canals, thereby contaminating the groundwater from the surface. Fig. 8.9 portrays the increase in intrusion from 1904 to 1959. Increased salinity required several well fields to be abandoned in the Miami-Fort Lauderdale area.

Remedial measures have included construction of saltwater barriers in the drainage canals and the establishment of water management areas which serve to pond fresh water for aquifer recharge.

8.4.3 Los Angeles, CA

Saltwater intrusion was noticed in the early 1930s as a result of water usage for agricultural, domestic and industrial purposes. A water injection barrier project has been implemented along a 11 km portion of the shoreline. A total of 94 injection wells cause a pressure "ridge" in the confined aquifers, thereby establishing a seaward gradient which prevents seawater encroachment. The piezometric head of the barrier is maintained at 1 to 3 m above sea level, requiring approximately 1,500 m^3/day of injected water. Figure 8.10 presents a cross section perpendicular to the shoreline which demonstrates the increase in the pressure ridge from 1963 to 1967.

8.4.4 The Potomac-Raritan-Magothy Aquifer System

Lennon et al. (1987) have reported on a numerical model study to evaluate the effectiveness of creating a hydraulic barrier to salinity inflow for the Potomac-Raritan-Magothy (PRM) aquifer system. The optimum

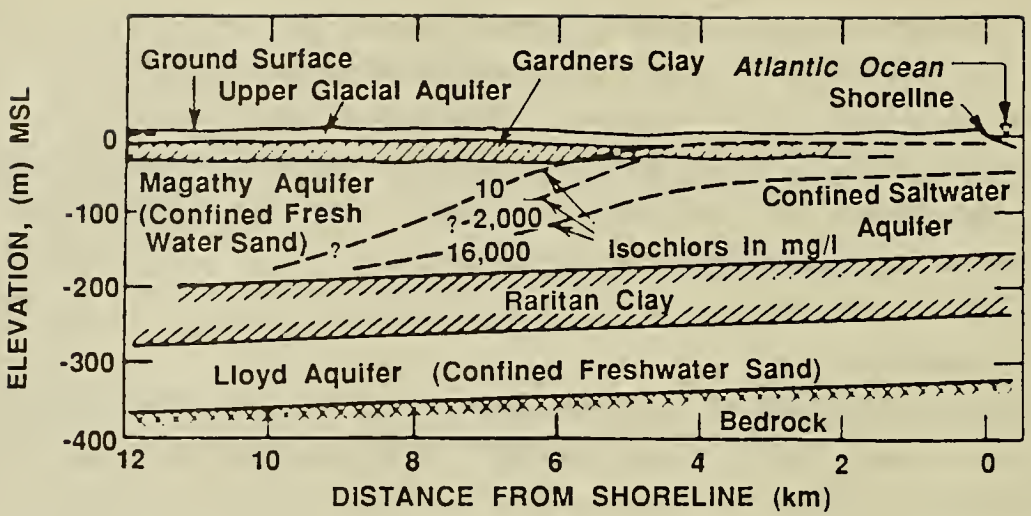


Fig. 8.8 Profile through Aquifer at Far Rockaway, Nassau County, Long Island, Showing Location of Salinity Front as a Result of Pumping (reprinted with permission from D. K. Todd, Groundwater Hydrology, copyright 1980, John Wiley & Sons, Inc.).

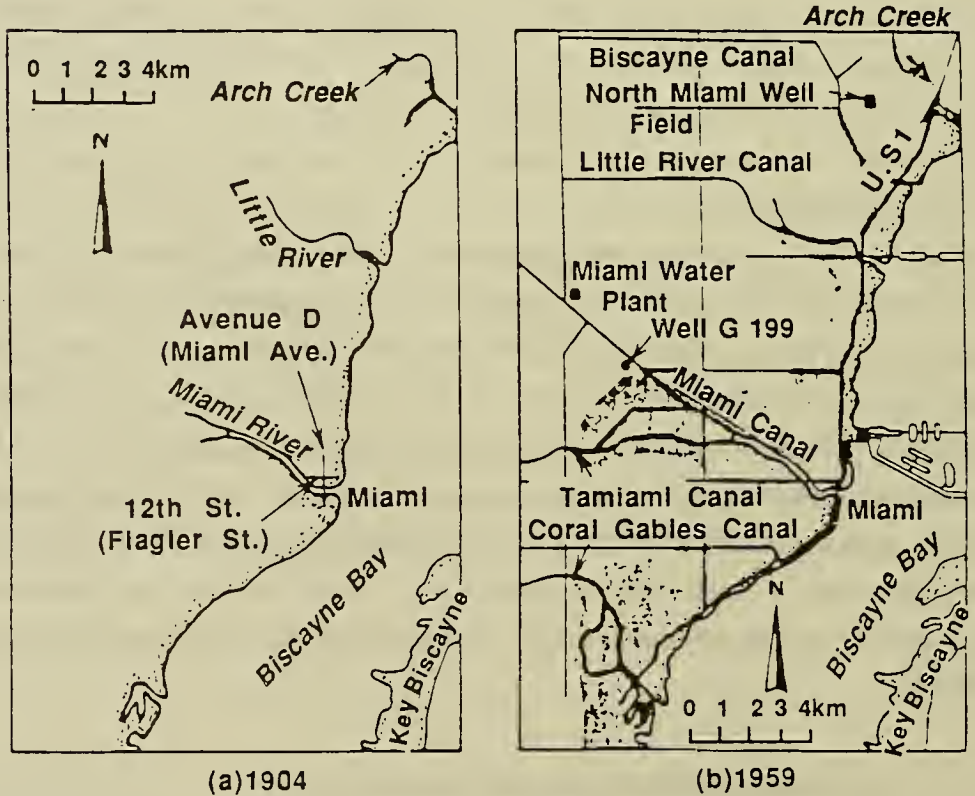


Fig. 8.9. Progressive Saltwater Intrusion in the Vicinity of Miami, FL, 1904 to 1959 (reprinted with permission from D. K. Todd, Groundwater Hydrology, copyright 1980, John Wiley & Sons, Inc.).

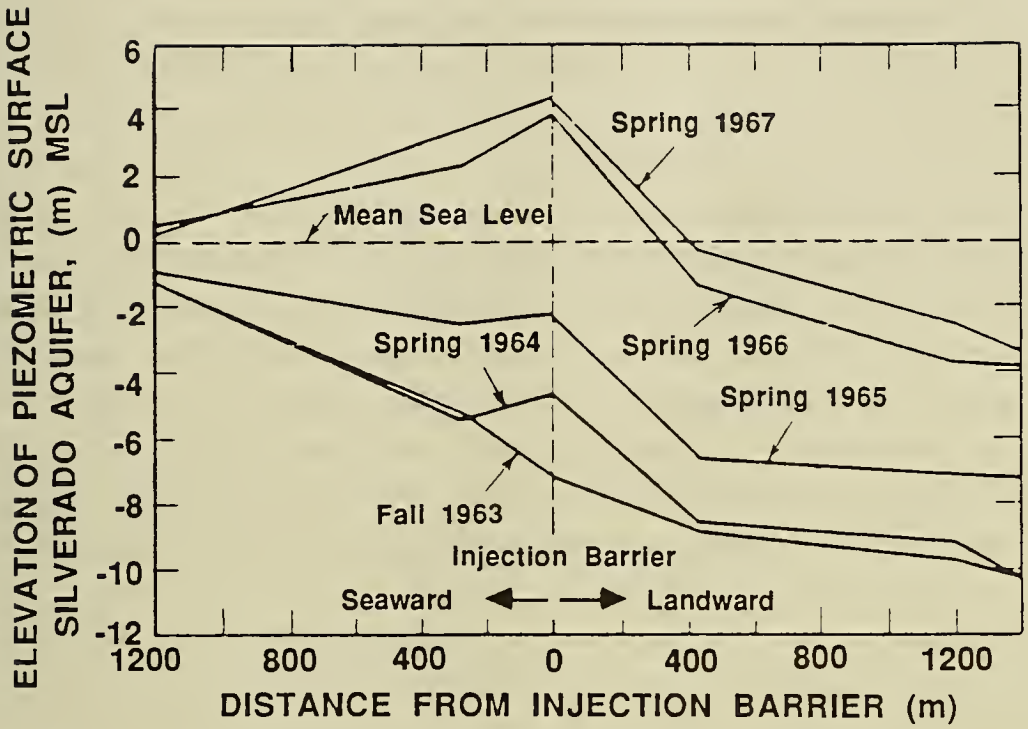


Fig. 8.10. Piezometric Pressure Profiles Perpendicular to the Seawater Intrusion Barrier in Los Angeles County for Various Times after Commencement of Injection in the Fall of 1963 (reprinted with permission from D. K. Todd, Groundwater Hydrology, copyright 1980, John Wiley & Sons, Inc.).

approach was a combination of extraction and injection wells. The aquifer length (parallel to the Delaware River) considered was 1,520 m and 5 injection wells were located parallel to and 300 m from the Delaware River and the 5 extraction wells were located 600 m from the river. Only 0.014 m^3/s was pumped from each extraction well and reintroduced into the aquifer through the adjacent injection well. This resulted in a freshwater "mound" that performs as a barrier to saltwater inflow. Following the drought, the injection wells are used as extraction wells (0.014 m^3/s) with the water returned to the river, thereby reducing the salinity in the aquifer further.

8.4.5 Okinawa-jima Island

The construction of underground flow barriers through pumping of cement grout near the coastline can create underground impoundments similar to surface reservoirs. In 1978, such an experimental subsurface barrier was constructed in a small buried valley on Miyako-jima Island (Sugio et al., 1987). The barrier was constructed in a very porous limestone aquifer and is 16.5 m high, 5 m wide and 500 m long. Based on field monitoring studies conducted over a four year period, the installation was judged a success; the hydraulic conductivity and porosity were reduced from 0.17 cm/s and 20% to 5×10^{-5} cm/s and 6%, respectively. Plans are under way to construct a much larger barrier at Komesu on Okinawa-jima Island where the annual rainfall is 2,400 mm and occurs during a nine month period. With the present high permeability, much of this fresh water flows to the sea and saltwater intrusion tends to occur during the remaining three months when heavy pumping is conducted for irrigation purposes.

Based on numerical modeling with a particular barrier design of 5 m thickness, it is concluded that with the barrier no salinity intrusion will occur for a 60 day period during the drought and that if a longer period is desired, the barrier thickness must be increased or the pumping rate (6,000 m^3/day) must be decreased.

8.5 RESEARCH NEEDS

Given the ambient flow conditions in a coastal aquifer, the transmissive and porosity properties of the aquifer and various scenarios

of extraction "demand" on the aquifer, it appears that the characteristics of the salinity intrusion, including any time dependencies can be predicted reasonably reliably by state-of-the-art numerical modeling. However, at present the effects of sea level rise on saltwater intrusion have been examined only for special cases. Any decisions by those responsible for planning related to the need to relocate well fields, modify usage, implement remedial measures, etc. must be based on realistic estimates of the effects of sea level rise and other causes such as increases in extraction rates.

A research program is recommended which would develop and exercise rather simple models specifically with the objective of illustrating the relative effects of sea level rise. The remaining efforts would be much more comprehensive and concentrate on evaluating the reliability of numerical models for predicting salinity intrusion and on case studies of areas where the potential for salinity intrusion is high. These studies would focus on the applications of models, parameterized and calibrated for the local aquifer characteristics and using various sea level rise scenarios to predict effects on extracted water and on the need for and effectiveness of various types of remedial measures.

9. UPRIVER SALTWATER PENETRATION

9.1 INTRODUCTION

An estuary by definition is a semi-enclosed water body in which seawater and fresh water from river mix under tidal action. Seawater is thus diluted measurably (Cameron and Pritchard, 1963), and, in some cases, penetrates in the form of a saline wedge upriver. In other cases the waters are vertically well mixed, and salt water penetration occurs without the presence of a distinct wedge. Water bodies which do not receive fresh water tend to be highly saline, with salt concentration equal to that in the sea. In some cases, e.g., in parts of Florida Bay, in southern Florida, excessive evaporation during the hot and dry season renders the waters super-saline, with salinity exceeding that in the sea (Atlantic Ocean).

The three main parameters which control the degree of salt penetration are the river runoff velocity, water depth and tidal range in the sea. Increasing the tidal range or the depth, or decreasing the runoff will increase penetration. In urbanized areas, withdrawal of fresh water and dredging of deeper channels for vessel navigation are important issues; hence the effect of reduced runoff as well as the effect of increased depth on salinity intrusion have been investigated by scientists and engineers. The effect of reduced runoff has, for example, been recently investigated in the Myakka and other rivers near Florida's Gulf of Mexico coast.[1] A serious problem of this nature occurred in the Delaware River Basin in the 1960s due to drought. During the worst period, the salt front advanced 53 km up the river and forced some industries near Philadelphia to seek water from a municipal system that imported water from the Susquehanna River Basin (Hull and Titus, 1986).

The influence of increased depth is analogous to what would occur (and has occurred) in the event of a sea level rise. The propagation of tide up the estuary is affected in this case. On the other hand, enhanced fresh

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[1]Ernest Esteves, Mote Marine Laboratory, Sarasota, Florida, personal communication.

water withdrawal only partially simulates the sea level rise effect, since reduced runoff does not influence tides as significantly as does an increased water depth.

9.2 LITERATURE REVIEW

The significance of salt water penetration has been a matter of common knowledge among scientists and engineers for a long time, but the entire process was placed on a firm physical footing relatively recently. Perhaps the most important work, at least in the United States, was carried out during the post-second world war period by Keulegan at the National Bureau of Standards, by Ippen at M.I.T., and later by Harleman also at M.I.T., in cooperation with the U.S. Army Engineer Waterways Experiment Station at Vicksburg, Mississippi. These works, summarized in Ippen (1966), were primarily laboratory-based, and sought to understand the intrusion process through the development of basic, analytic formulations. It is noteworthy that, in a sense, these investigations were successors to the pioneering work of O'Brien (O'Brien and Cherno, 1934), which dealt with the fundamentally similar problem of predicting the rate at which salt water intrudes into fresh water, as in a lock separating a saline water body from one of lower salinity. Partheniades (1971) has reviewed the fundamentals of the salt water intrusion mechanisms (see also Partheniades et al., 1975). Beginning in 1954, the Committee on Tidal Hydraulics of the Corps of Engineers has issued a series of reports pertaining to various theoretical and practical aspects of the estuarine salinity intrusion problem.

Concurrently with the aforementioned basic laboratory-analytic studies, physical models of real estuaries were developed, with the inclusion of saltwater intrusion effects. The Corps of Engineers led this effort, and constructed a rather large model of the San Francisco Bay estuarine system in Sausalito, California (Fischer et al., 1979). This is a distorted model (scales 1000 horizontally and 100 vertically), i.e., one in which the water depths are greater than the linear scale set by the aerial (horizontal) dimensions. Distortion is commonly employed in physical models in order to generate sufficient turbulence necessary to satisfy the desired equivalence between model and prototype flow

conditions. Distortion, however, essentially requires additional flow resistance to be provided at the bottom by means of artificial roughness elements. Calibration of such a model, particularly one involving two fluids, can be a tedious process. It took several years to make the San Francisco Bay model fully operational. Other major physical models of this nature are those of the Mississippi River and Chesapeake Bay.

Extensive field investigations were carried out in the fifties for understanding salinity intrusion through direct (prototype) evidence, and to calibrate physical models such as the ones noted above. Reference may be made to the investigation by Sir Claude Inglis (Inglis and Allen, 1957) on the Thames in England, and by Prof. Pritchard (Pritchard, 1952) on the Chesapeake Bay. In both estuaries, as elsewhere, studies which began in the fifties (and earlier) have been of an ongoing nature, and have continued until the present time. The reason for this is both a continued interest in the basic aspects of the mechanism of mixing between salt water and fresh water in the real environment (see, for example, Dronkers and van de Kreeke, 1986), and also because new engineering problems continually arise and, therefore, must be examined afresh. Physical models of large estuarine systems such as the San Francisco Bay, Chesapeake Bay, the Mississippi and New York Harbor have been retained, and are used as needed by the Corps of Engineers and other agencies. As fresh input for modification and calibration of such models, prototype studies are conducted, although many now tend to be highly site-specific, given the costs involved in field work.

Computer technology has made it possible to develop sophisticated numerical models for handling estuarine hydrodynamics including salinity intrusion. The early models, in the sixties and early seventies, were typically one-dimensional, simulating cross-sectional average processes in the longitudinal direction (e.g., Harleman et al., 1974; Miles, 1977). These were followed by two-dimensional, depth-averaged models, e.g., such as the one incorporated in the TABS-2 system of estuarine numerical models used by the Waterways Experiment Station (Thomas and McAnally, 1985). More

recently, fully three-dimensional models have been developed. These models have been applied to a study of New York Harbor[2].

9.3 PHYSICAL PRINCIPLES

Analytic approaches to solve the problem of saltwater intrusion depend upon whether the estuary can be treated as stratified, or as partially mixed or fully mixed. Thus, for example, an estuary is classified as "well mixed" if the salinity over any flow cross section does not vary by more than 10%; otherwise it is classified as partially mixed (Ippen, 1966). In a fully stratified estuary the salt is contained almost entirely in the lower layer (the wedge). A further quantitative means of classifying and comparing estuaries, and one which requires salinity and velocity only, is due to Hansen and Rattray (1966). Their commonly used diagram relates, for different types of estuaries, a stratification parameter, $\delta s/s_o$, defined as the ratio of the surface to bottom difference in salinity, δs, divided by the mean cross-sectional salinity, s_o, and a circulation parameter, u_s/u_f, the ratio of net surface current, u_s to the mean cross-sectional velocity, u_f.

The penetration of salt water into fresh is strongly influenced by density differences, since salt water is the heavier of the two fluids. Consider a barrier which separates two fluids, one of density ρ (water) with zero salinity ($s = 0$) and another of density $\rho + \rho\Delta$ (salt water), salinity s_o. As shown in Fig. 9.1a, if the barrier were lifted at time $t = 0$, a saltwater gravity front (current) would penetrate fresh water at the bottom, and an equal volume of fresh water would be displaced per unit time into salt water, at the top. If now a freshwater flow of velocity, U_o, were imposed as shown in Fig. 9.1b, the intrusion of gravity current would be arrested at some position where the pressure and drag forces in both directions balance. This is the basic mechanism which operates within estuaries, where sea-driven tides are an additional factor. In cases where the mixing potential of the incoming tidal energy is relatively high, the estuary will be well mixed in terms of vertical salinity structure. If

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[2]Allen Teeter, Waterways Experiment Station, Vicksburg, Mississippi, personal communication.

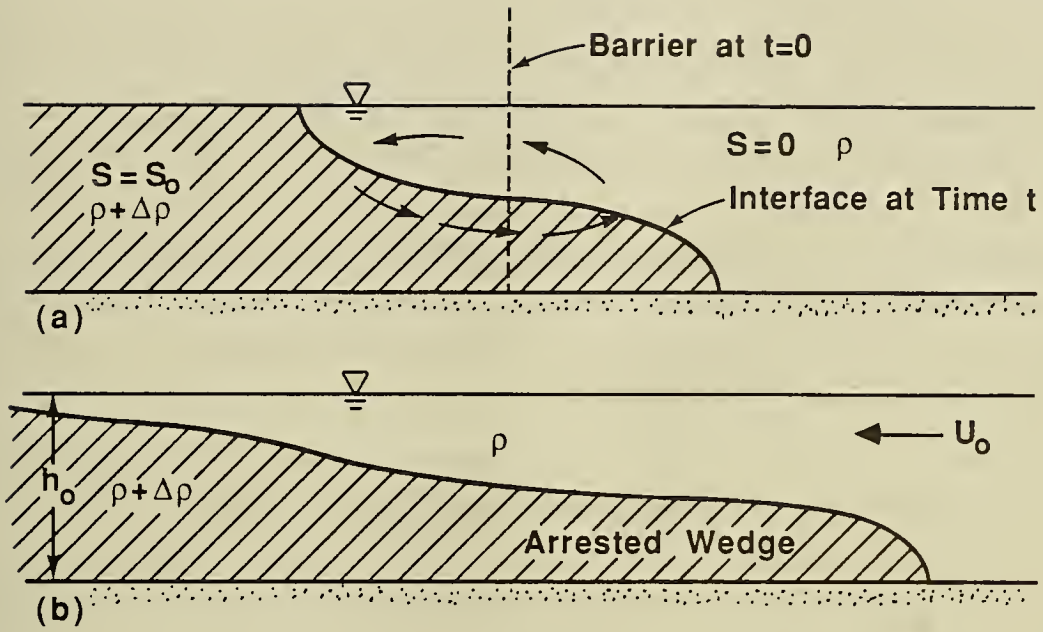

Fig. 9.1. Mechanism of Salt Penetration: a) Development of a Gravity Current, b) Arrested Saline Wedge.

not, the estuary will be stratified. Since both the tidal range and runoff vary; the former principally on a synodic basis while the latter seasonally, many estuaries shift between fully stratified and well-mixed conditions over a year period (Dyer, 1973). In some cases, very high runoff can virtually "clean out" the estuary of salt water (e.g., during and immediately following storm runoff or during the monsoon season) (Rao et al., 1976; Mehta and Hayter, 1981).

During a tidal cycle, the water level rises and falls, and this variation directly influences the length of penetration of salt water; at high water, salt water penetrates further than at low water. It is instructive to look at the basic governing equations and their behavior in this context.

Ippen and Harleman (Ippen, 1966) considered the basic unsteady, one-dimensional (x,t) mass conservation for salt transport-diffusion, and derived the following expression for the resultant profile of salinity, s(x,t) in a non-stratified estuary:

$$\frac{s(x,t)}{s_o} = \exp\left\{-\frac{U_o}{2D_o'B}\left[N-(N-x)\exp\left(\frac{a_o}{h_o}(1-\cos\sigma t)\right)+B\right]^2\right\} \qquad (9.1)$$

where $N = h_o u_o / a_o \sigma$. Here, U_o = freshwater outflow velocity, D_o' = diffusion coefficient, B = an empirical, diffusion-related coefficient, a_o = tidal amplitude, σ = tidal frequency, u_o = maximum tidal velocity at the mouth and h_o = mean water depth.

If the end of the intrusion zone is specified at $s/s_o = 0.01$, the maximum intrusion length, L_{max}, occurs when $\sigma t = \pi$, and the minimum, L_{min}, when $\sigma t = 0$. Thus

$$L_{max} = N\left[1-\exp\left(-\frac{2a_o}{h_o}\right)\right] + B\left(3\frac{D_o'}{U_o B}-1\right)\exp\left(-\frac{2a_o}{h_o}\right) \qquad (9.2)$$

$$L_{min} = B\left(3\frac{D_o'}{U_o B}-1\right) \qquad (9.3)$$

In the above equations, the parameters D_o' and B must be known. In a real estuary, these parameters can be determined by measuring the average salinities at two points in the estuary at low tide.

Examples of average salinity distribution at high and low water slack as shown in Figs. 9.2a,b are based on the experiments in a model tidal channel at the Waterways Experiment Station (Harleman and Abraham, 1966). This illustration of the difference in salinity intrusion between high water slack and low water slack (tidal range 3 cm) is essentially analogous to what would occur under an equivalent sea level rise. Note that Figs. 9.2a,b clearly show that salinity intrusion is a highly dynamic phenomenon, and that there is a 21.4 m difference in the distance of penetration between the high and low water events. It is thus evident that only a small head is required to cause a significant horizontal movement of the salt water front. Raising the sea level (increasing h_o in Eq. 9.1) would amount to pushing both curves up the estuary. The same would occur if h_o were increased by dredging. Likewise, decreasing U_o, the outflow velocity, would cause further penetration. This can be easily shown via Eq. 9.2, for example. A possible secondary effect in the event of a sea level rise could be higher tides at the mouth. In this case a_o would increase, thereby reinforcing the penetration effect produced by increasing h_o.

The aforementioned simple, demonstrative theory of course has major limitations by virtue of the assumptions under which the stated equations were derived. In general, however, many of the restrictions can be relaxed via solution of the fully three-dimensional transport equations numerically. Furthermore, by following a hybrid approach involving a combination of field data, physical model-based data and numerical simulation, it is presently feasible to arrive at realistic descriptions of salinity distributions in the estuary, and such effects as those due to sea level rise can be simulated with a reasonable degree of confidence, particularly in the tide-dominated (as opposed to waves) environment. It should, however, be noted that the costs involved in "full blown" studies remain high, with the result that salinity intrusion effects have been investigated so far, in detail, mainly in highly urbanized estuaries.

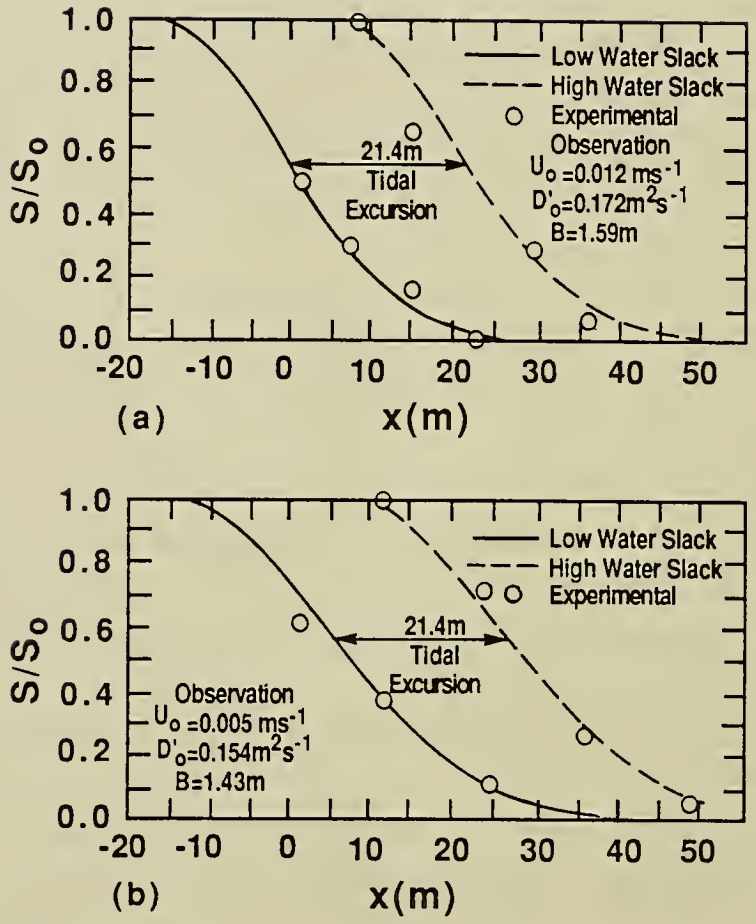

Fig. 9.2. Longitudinal Salinity Distribution in a Model Tidal Channel: a) Test No. 2, b) Test No. 16 (after Harleman and Abraham, 1966).

9.4 EXAMPLES

It suffices to examine the effects of channel deepening and runoff on salinity intrusion as paradigms (qualitative in the case of runoff) for what would occur in the case of a sea level rise.

A major salinity intrusion problem developed in the Maracaibo estuary, Venezuela, in the sixties (Partheniades, 1966). The problem was traced to the construction, during the previous years, of a deep, 60 km long, navigation channel connecting Lake Maracaibo to the Gulf of Venezuela. A representative salinity (chlorinity) record within the lake during the 1937-63 period is shown in Fig. 9.3. A significant fluctuation of salinity first occurred during the 1947-52 period. This rise in salinity, from typical values of 500-700 ppm during the previous and the subsequent years, to a peak of about 1,400 ppm, is believed to have been due to the occurrence of a dry period with low runoff. However, following the completion of the deeper (from 6 m to 11 m) channel in 1956, the salinity increased relatively rapidly and almost continuously to a peak of 1,500 ppm in 1959. At this time, to accommodate yet larger oil tankers, the channel was further deepened to 13.5 m. This caused a further increase in salinity and, in 1962-63, when the deepening project was completed, the salinity rose to 2,200 ppm. A physical model of the estuary was subsequently constructed to examine the problem in further detail (Brezina, 1975). The ultimate outcome of the channel deepening projects was to turn Lake Maracaibo from a relatively fresh water lake into a brackish water lagoon, with a complete change in life forms[3].

The effect of dredging is commonly evaluated for environmental impact statements. Such a study was carried out as prerequisite to the construction of the Trident submarine base at King's Bay, Georgia. The ocean entrance to this bay is through St. Mary's Entrance, Florida. Fig. 9.4 shows the application of Eq. 9.1 for predicting the high and low water (tidal range 2.2 m) salinity profiles in this prototype case, prior to dredging (Parchure, 1982). Although agreement between theoretical curves and measured data is obviously very rough, the theory, although

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[3]Jindrich Brezina, University of Zulia, Maracaibo, Venezuela, personal communication.

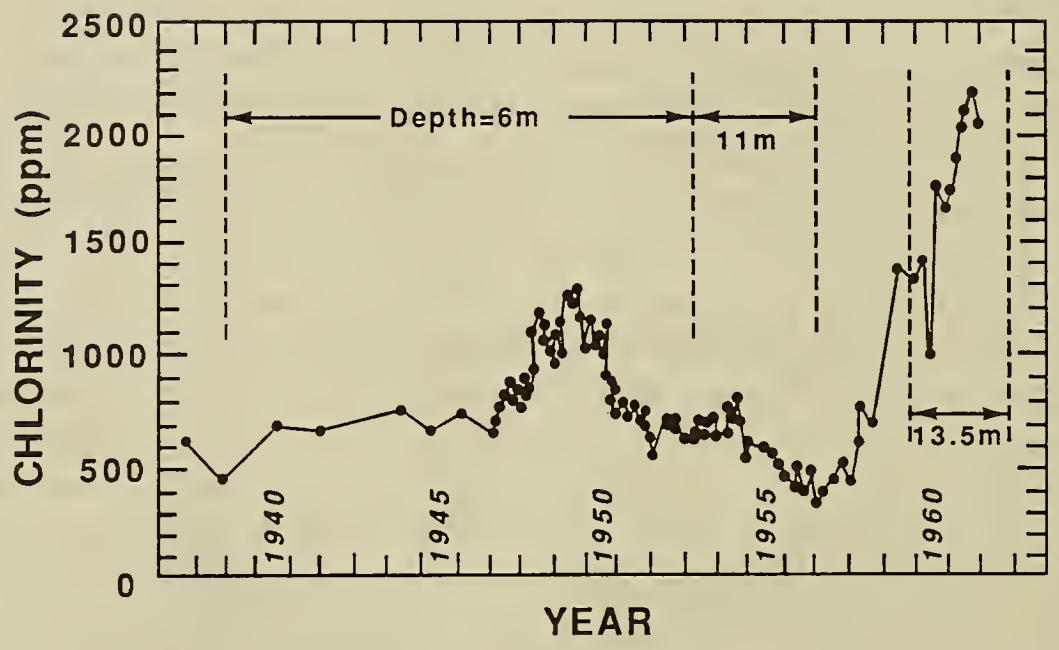

Fig. 9.3. Salinity (Chlorinity) Variation with Years in Lake Maracaibo (after Partheniades, 1966).

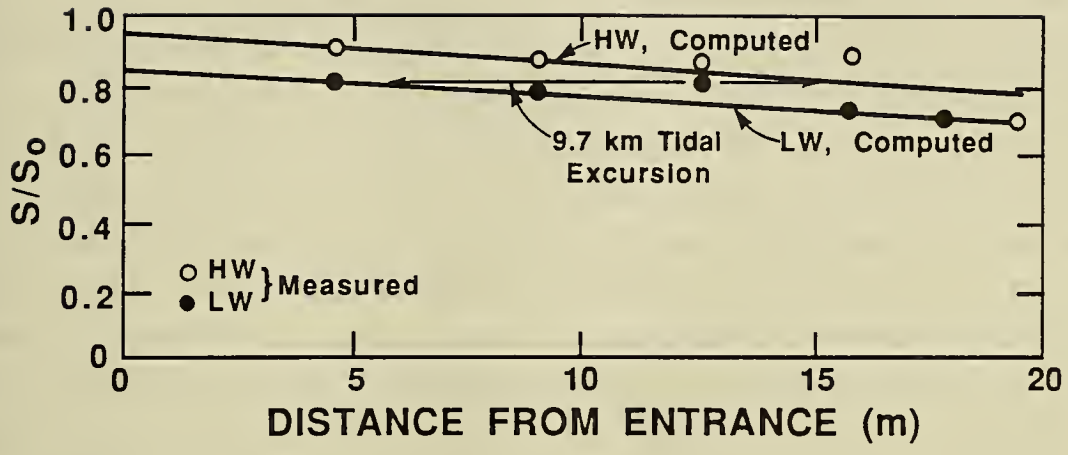

Fig. 9.4. High and Low Water Salinity Profiles through St. Marys Entrance, Florida and Cumberland Sound, Georgia (after Parchure, 1982).

approximate, does simulate the overall trends suggested by the measurement. Dredging plans called for a deepening of the channel between the entrance (end of jetties) and King's Bay ranging from 0 to 4 m. A 4% rise in the salinity was predicted 20 km up the entrance as a result of dredging (Environmental Science and Engineering, Inc., 1980).

The aforementioned examples are merely illustrative of the basic phenomenon of interest, and are not meant to demonstrate the power of available technology with respect to physical or numerical models and their application. The subject matter has been covered extensively in literature on fluid mechanics and hydraulic engineering. For a fuller treatment, the work of Fischer et al. (1979) may be cited. Here, it will suffice to make reference to a recent study conducted to examine the potential effects of deepening the lower Mississippi River on salinity intrusion (Johnson et al., 1987). In Fig. 9.5, the duration of the saline wedge intrusion (days per year) is plotted against distance above the head of a number of passes (distributaries) in the vicinity of New Orleans. Two curves are shown, one for a 12 m deep channel (present depth is 9 m) and another for a 17 m deep channel. These curves were generated numerically. The model was calibrated against data obtained during the 1953-54 low-flow period. The results essentially are illustrative of the effect of channel deepening from 12 m to 17 m. Thus, for example, this effect would increase the duration of the wedge at 180 km from about 40 days to about 60 days per year.

9.5 RESEARCH NEEDS

Scientific work in understanding the basic processes of mixing between salt and fresh waters is likely to continue well into the future, in order to steadily improve predictive methodologies, which have at present attained a reasonable degree of sophistication. However, further improvements are highly desirable.

An area which deserves further consideration is mixing under wave action, or combined wave and tide action. At present the effect of waves in this respect is only weakly understood. Much of analysis carried out so far appears to have been directed towards situations dominated by tides

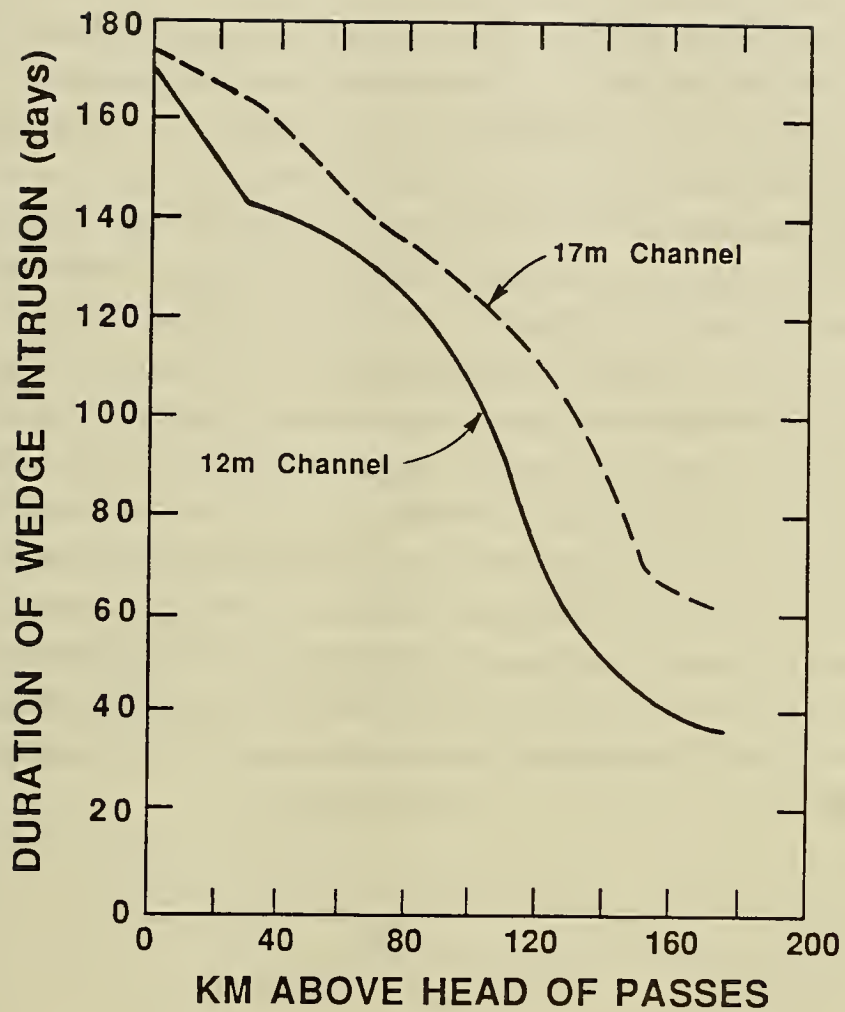

Fig. 9.5. Effect of Channel Deepening on the Duration of Wedge Intrusion in the Lower Mississippi River (after Johnson et al., 1987).

alone. It is noteworthy as well that with deeper water associated with sea level rise, waves would be admitted more freely, thereby decreasing fluid stratification by virtue of greater mixing in the vertical direction, particularly in estuary mouths.

Another important research area is related to the development and motion of fronts due to salinity, temperature and sediment density gradients. Understanding the behavior of these fronts is vitally important to a range of water quality and ecological issues.

It must be mentioned that most of the work done to date appears to have dealt with such effects as those related to channel deepening or changes in upstream river hydrology, rather than sea level rise. As noted, sea level rise can also increase the coastal tide and wave action, particularly if the coastline is rocky, and does not recede, while water depth increases. A similar situation can also arise if the shoreline is erodible, but sea level rise is so rapid as to prevent depths at the mouth from achieving quasi-equilibrium with the hydrodynamic forcing. This would lead to a situation wherein the estuarine mouth would be in deeper water, and where the tide (and waves) would arrive with lesser hindrance due to reduced bottom friction and lesser chance of wave breaking in the offshore waters. In most work carried out so far, the coastal tide (and waves) is typically assumed to practically remain unchanged.

10. SEDIMENTARY PROCESSES IN THE ESTUARINE REGION

10.1 INTRODUCTION

There have been drastic changes in the world's shorelines since the retreat of the last ice age. Over the past ~6,000 years, the rate of rise of sea level has been relatively low (0.08 m/100 yr) compared to the period ~20,000-6,000 BP (0.8 m/100 yr) (Fig 7.1). It is not surprising, therefore, that many of today's estuaries have been "around" approximately in their present configuration only in the past few millennia. Even in the absence of sea level change, estuaries are highly dynamic and in many ways reflect the type of macro-scale processes characteristic of oceans. Estuarine shorelines change under the action of hydrodynamic forcing and associated sediment transport. Where sediments play a recognizable role, estuaries almost never attain true hydrodynamic/sedimentary equilibrium. Usually there is a quasi-equilibrium characterized by long-term changes in the bottom bathymetry. An important issue posed by potential effects of sea level rise pertains to our ability to predict various facets of estuarine response to sea level rise, including shoreline configuration, bottom sedimentation and marsh development/degradation.

10.2 SHORELINE CONFIGURATION

Much of the knowledge of shoreline changes is based on geological evidence which has been used to develop scenarios for estuarine formation, development and eventual demise. Two edited volumes by Schwartz, Spits and Bars (1972) and Barrier Islands (1973), are collections of important papers in the subject area. Spits and Bars covers seventeen papers, from 1890 to 1971. Barrier Islands covers forty papers, from 1845 to 1972.

A more recent set of papers (also of geological nature) edited by Leatherman (1979) shows that there is new emphasis on descriptive modeling of barrier island and inlet morphologic changes, much of it based on holocene shoreline recession evidence. Reference must also be made to a series of papers recently edited by Nummedal et al. (1987) on shoreline response to sea level change.

Our understanding of processes at a particular site, post facto, has improved vastly. However, prediction of future shorelines, in any given situation, at best remains a hazardous guess except in a few well defined situations. In many cases therefore, simple modeling based on inundation and the Bruun Rule are used for predicting long-term trends (Kana et al., 1984). The problem seems to lie with the fact that our ability to predict sediment motion along the coastline under a changing wave climate on a long-term basis remains rudimentary, and is in fact the subject of major ongoing research effort in coastal engineering. It may take an additional decade or two before confidence in long-term prediction reaches acceptable levels. A part of the difficulty is not so much with relating hydrodynamic forcing to sediment motion, as with synoptic data necessary to obtain reliable (hindcast) hydrodynamic (currents, waves, winds) information. Furthermore, structures interrupt sediment motion. Modeling of sediment-structure interaction is still under research, although some useful modeling work has been done in this context (Kraus, 1983).

In the subject of estuarine mouth or inlet response to changing oceanographic conditions, much more work has been done on sandy inlets than on inlets where the material is fine-grained. In general, simplified description of inlet response can be examined by considering inlet hydraulics as characterized by the repletion coefficient concept (Keulegan, 1967) and inlet sedimentary response via O'Brien's (1969) equilibrium relationship for inlet stability.

The repletion coefficient, K, is defined as

$$K = \frac{A_c}{a_o \sigma A_b} \left(\frac{2ga_o}{F}\right)^{1/2} \tag{10.1}$$

where A_c = inlet flow area, A_b = bay area, σ = tidal frequency, a_o = tidal amplitude in the sea, g = acceleration due to gravity and F = impedance. F accounts for bottom resistance in the channel as well as head losses associated with flow entrance and exit (Keulegan, 1967; Mehta and Özsoy, 1978). Sea level rise will influence several terms, including A_c (increase), A_b (increase, unless bay is bounded by vertical walls), a_o

(assume increase) and F (decrease). Typically, the overall effect will be an increase in K, which means easier flow admittance or better repletion.

The O'Brien relationship between the spring tidal prism, P, and A_c is

$$A_c = aP^m \tag{10.2}$$

where a and m are empirical coefficients which vary somewhat with the prevailing wave climate and other local conditions. The significance of this relationship is that it implies inlet widening and/or deepening with prism. Sea level rise in most cases will increase the prism (greater repletion), to which the inlet flow area will respond likewise. Equation 10.2 is particularly applicable to sandy inlets, as described in section 6. Increasing prism means greater sand flushing ability of the channel. The sand is transported by higher currents out of the channel, both bayward and seaward, to flood and ebb shoals, respectively. With increasing prism, there is likely to be a corresponding increase in the volume of these shoals. Furthermore, as the sea level rises the deltas must grow in elevation to keep up with the rise, implying that any natural bypassing of sand would reduce and that downdrift erosion would increase.

Stabilized inlets will be affected strongly by a large sea level rise. The protective jetties, which retard the ability of the littoral drift to enter the navigational channel and reduce the wave climate in the channel, would become less effective as they are submerged. Also, the stability of the jetties may be reduced due to increased wave heights as a result of sea level rise (National Research Council, 1987b).

The sea level rise scenario imposes a much more gradual change in the inlet/bay system than, for example, channel dredging. There should be enough time for the system to keep pace with water level rise, with the attainment of quasi-equilibrium under the prevailing hydrodynamic forces and sediment movement. Equilibrium is determined jointly by hydraulic conditions characterized by the repletion coefficient (Eq. 10.1) and by sedimentary requirement as per Eq. 10.2 (O'Brien and Dean, 1972).

It is also self-evident that shoreline response as far as inlets are concerned is contingent upon the availability or lack of sediment supply. For example, the barrier islands of the Mississippi-Alabama coast have been

migrating and "disappearing" due to lack of sediment supply (Otvos, 1979). See also Fig. 6.1. Over the past ~100 years, this factor seems to have been far more important than (absolute) sea level rise.

The East Frisian Islands in Germany illustrate trends of an opposite nature. These islands and the corresponding inlets are believed to have been in existence for the past few millennia (Luck, 1976). Unlike the Mississippi barriers, however the islands have accumulated additional sediment from the West Frisian Islands. Fig. 10.1 shows three of these islands (and four inlets) [Norderney (N), Baltrum (B), and Langeoog (L)], plus ends of two [Juist (J) and Spiekeroog (S)]. It is observed that between 1750 and 1960, the barrier as a whole gained sediment, resulting in greater land area above mean high water (MHW). To some extent, these developments have been influenced by coastal protection works at the eastern ends of Norderney and Baltrum (Kunz, 1987). Luck (1976) has accumulated map-based evidence on the changes at East Frisian Islands and has provided a qualitative explanation for the observed historic behavior. Unfortunately, present day technology precludes the possibility of predicting changes in island chain configuration over even the next 10 years.

10.3 ESTUARINE SEDIMENTATION

It is common for inlets to have sandy beds and upland estuarine waters to consist of fine-grained material. A relatively rapid sea level rise could cause a correspondingly rapid landward migration of the sandy barrier island, thus exposing back barrier fine-grained deposits to open coastal wave action (Everts, 1987). In this way, a marine source of fine-grained sediment is created. Some low coasts lack barriers, for example west Florida north of Tarpon Springs. Barriers have not developed here because of the lack of sandy material, which in turn is due to the limestone formations through which the coastal rivers drain (Shepard and Wanless, 1971). Here again there is an offshore source of fine-grained sediment which generates coastal turbidity mainly during storms.

The significance of a marine source of fine-grained sediment is that this material tends to enter the estuary with the upstream salinity-driven residual current, which typically occurs in the mixing zone of the estuary

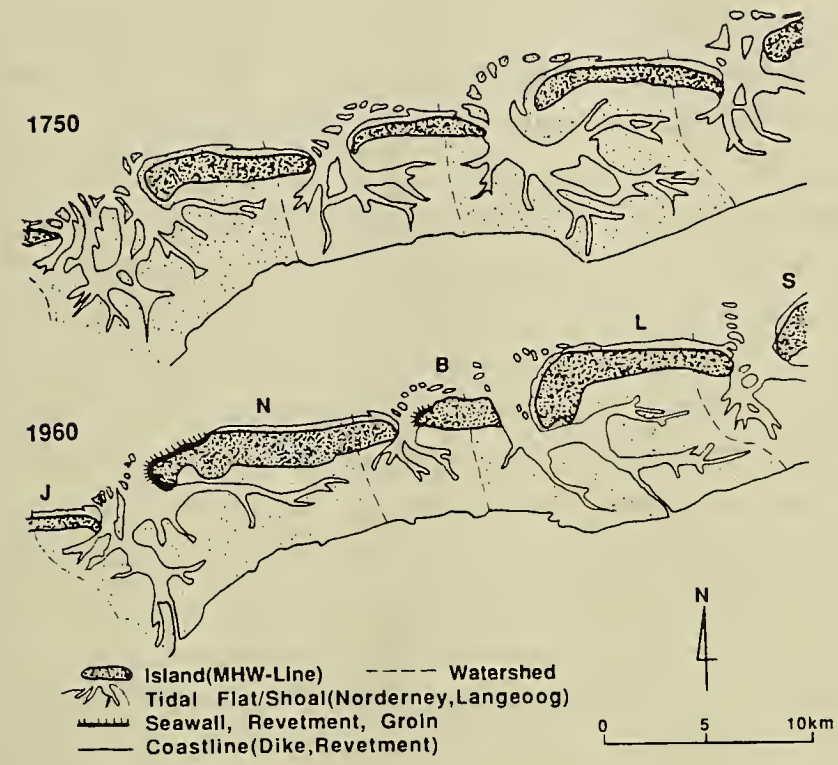

Fig. 10.1. East Frisian Islands in 1750 and in 1960 (after Kunz, 1987).

(i.e., the region where sea water mixes with fresh water). With sea level rise and barrier migration, fine-grained sediment from the sea would become available as a sediment source, stirred up by wave action and transported by the residual current, depositing ultimately at the upland end of the mixing zone. In some cases, e.g., in the long navigation channel through the Maracaibo estuary in Venezuela (Partheniades, 1971) and in Zeebrugge harbor, Belgium,[4] the present source of fine sediment is marine rather than alluvial.

Fine sediment transport in estuaries is a complex process involving a strong coupling between tides, freshwater flow and the coagulated sediment. This process has been described extensively elsewhere (Postma, 1967; Krone, 1972). In Fig. 10.2, a schematic description is given. The case considered is one in which the estuary is stratified, and a stationary saline wedge is formed. Various phases of suspended fine sediment transport are shown on a tide-averaged basis. In a partially mixed estuary, the description must be modified, but since relatively steep vertical density gradients are usually present even in this case, the sediment transport processes are qualitatively similar.

With reference to this figure, the vertical variation of the horizontal residual flows (tide-averaged flows) can be conveniently described by computing the ebb predominance factor, EPF, defined as

$$EPF = \frac{\int_0^{T_E} u(z,t)dt}{\int_0^{T_E} u(z,t)dt + \int_{T_e}^{T_F} u(z,t)dt} \qquad (10.3)$$

where $u(z,t)$ = instantaneous longitudinal current velocity at an elevation z above the bed, T_E = ebb period, T_F = flood period and $T = T_E + T_F$, where T = tidal period (Simmons, 1965). If the strengths of flood and ebb were the same throughout the water column, EPF would be equal to 0.5 over the entire depth of flow. This is almost never the case; EPF is usually less

[4]Robert Kirby, Ravensrodd Consultants Ltd., Taunton, United Kingdom, personal communication.

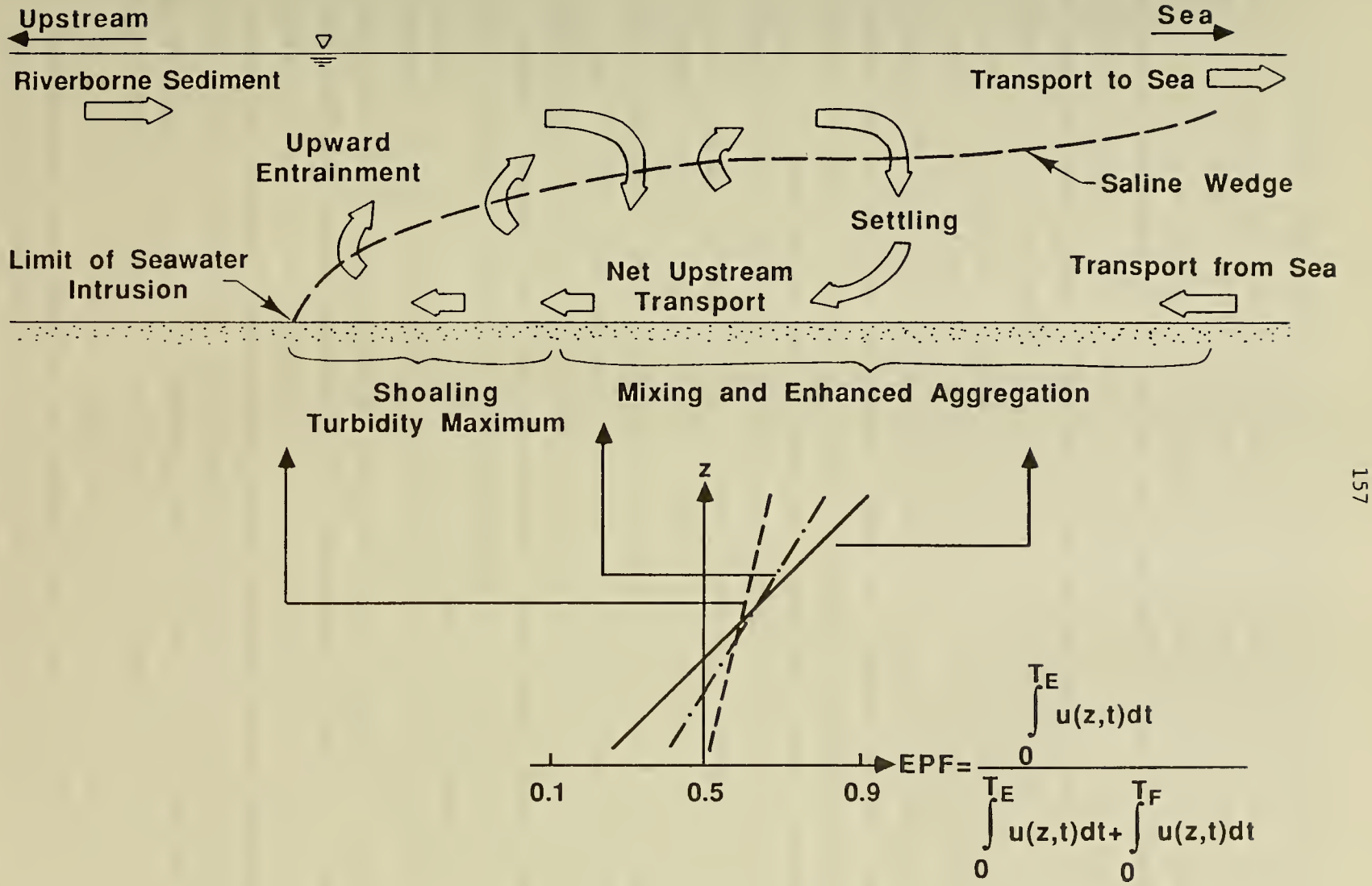

Fig. 10.2. Sediment Transport in the Estuarine Mixing Zone.

than 0.5 near the bottom, particularly in the saline wedge, and greater than 0.5 in the upper layers. The residual upstream bottom current is due to the characteristic nature of flow circulation induced by the presence of the wedge, which means that the strength of this current will decrease as the limit of seawater intrusion is approached, and is theoretically zero at the limit (node) itself. Distributions of EPF at three locations - at the mouth, in the wedge and at the node, would qualitatively appear as shown in Fig. 10.2. When interpreted in terms of tidal flows, these distibutions reflect the general observation that in the mixing zone of the estuary flood flows landward at the bottom and ebb flows seaward at the surface.

The trends indicated by the EPF distibutions suggest the dominating influence of hydrodynamics on sediment movement. As noted in Fig. 10.2, riverborne (alluvial) sediments from upstream fresh water sources arrive in suspension in the mixing zone. The comparatively high degree of turbulence, associated shearing rates and the increasingly saline waters will cause sediment aggregates to grow in size as a result of frequent inter-particle collisions and cohesion, and large aggregates will settle. Aggregate settling velocities can be up to four orders of magnitude larger than the settling velocities of the elementary particles. Some of the sediment will deposit onto the bed, and some will be carried upstream near the bottom until times close to slack water when the bed shear stresses decrease sufficiently to permit deposition. The deposited sediment will start to consolidate due to overburden.

The depth to which the new deposit scours when the currents increase after slack will depend on the bed shear stresses imposed by the flow and the shear strength of the deposit. If the currents during both flood and ebb are sufficient to scour all of the new deposit, the net movement will be determined approximately by current predominance. However, if the bed shear stress during ebb is less than sufficient to suspend all of the newly deposited material, a portion of the material will remain on the bed during ebb, and will be resuspended and transported during the predominant flood flows, resulting in a net upstream transport. Net deposition, i.e., shoaling, will occur when the bed shear during flood, as well as during ebb, is insufficient to resuspend all of the material deposited during preceding slack periods. Some of the fine material that is resuspended

will be re-entrained throughout most of the length of the mixing zone to levels above the salt water-fresh water interface and will be transported downstream to form larger aggregates once again, and these will settle as before. At the seaward end some material may be transported out of the system. A portion or all of this could ultimately return with the net upstream current.

Sediment moving upstream along the bottom may also be derived from marine sources, as noted. The strength of this upstream current is often enhanced by the inequality between the flood and the ebb flows induced by the usually observed distortion of the tidal wave. Inasmuch as the low water depth is often significantly less than the depth at high water, the speed of the propagating tidal wave is higher at high water than at low water. This typically results in a higher peak flood velocity than peak ebb velocity and a shorter flood period than ebb period. Such a situation tends to enhance the strength of the upstream bottom current, and the sediment is sometimes transported to regions upstream of the limit of seawater intrusion.

The estuarine sedimentary regime is characterized by several periodic (or quasi-periodic) time scales. These are: 1) the tidal period (diurnal, semi-diurnal, or mixed), 2) lunar (spring-neap) cycle, 3) yearly cycle, and 4) periods greater than a year. Of these, the first is the fundamental period which characterizes the basic mode of the sediment transport phenomenon in an estuary. The second is important from the point of view of determining net shoaling rates in many cases of engineering interest, and by the same token the third and the fourth time-scales are involved in considerations of long-term stability and shoaling in estuaries, as for example due to sea level rise.

Predictive capability for estuarine sedimentation can be illustrated by considering two case studies, Atchafalaya Bay, Louisiana, and Savannah River estuary, Georgia. The Atchafalaya River, a distributary of the Mississippi River, discharges into this bay. In recent years, the delta at the mouth of the river has grown dramatically. A study of the bay and adjacent waters was carried out to predict the rate at which the delta will evolve in the short term (<10 years) and the long term (50 years), and the manner by which that evolution will affect flood stages, navigation channel

shoaling, and the environmental resources of the area (McAnally et al., 1985).

Several factors combined to make the Atchafalaya Bay study unusually complex. They included the long period over which predictions had to be made; the migration of the region of delta growth from lacustrine to estuarine to marine environments; a hydrodynamic regime that is variously dominated by river flows, wind-induced currents, tides, waves and storm surges; and the combined deposition of sediments from the sand, silt and clay classes. The investigation included several separate prediction techniques, including: 1) extrapolation of observed bathymetric changes into the future, 2) a "generic" analysis that predicted future delta growth by constructing an analogy between the Atchafalaya delta and other deltas in similar environments, 3) quasi two-dimensional numerical modeling of hydrodynamics and sedimentation, and 4) use of extensive field measurement of water levels, currents and sediments, and laboratory experiments on sediment samples.

Results have shown a wide range of possible land growth rates for the next 50 years in Atchafalaya Bay. Important results illustrated in Fig. 10.3 indicate sensitivity of delta growth to the subsidence rate (McAnally et al., 1984). These high subsidence rates are caused in part by compaction of thick layers of fine sediments that have been deposited by the Mississippi River and its distributaries over thousands of years.

The extrapolation results shown in Fig. 10.3 were generated by establishing a relationship between past delta growth and forcing phenomena of river flow and sediment supply, then using that relationship in combination with historically recorded flows to project future delta growth. It did not explicitly include subsidence effects. The generic approach assumed that Atchafalaya delta growth could be considered analogous to other deltas growing under similar conditions and matched observed Atchafalaya delta growth with a generic delta growth and decay time history such as shown in Fig. 10.3. The non-dimensionalized generic curve was then taken to represent a possible future growth and decay cycle for the Atchafalaya. Two results are shown--generic least squares fit the generic curve to observed Atchafalaya growth in a least squares approach; generic observed forced the generic curve to pass through the observed 1980

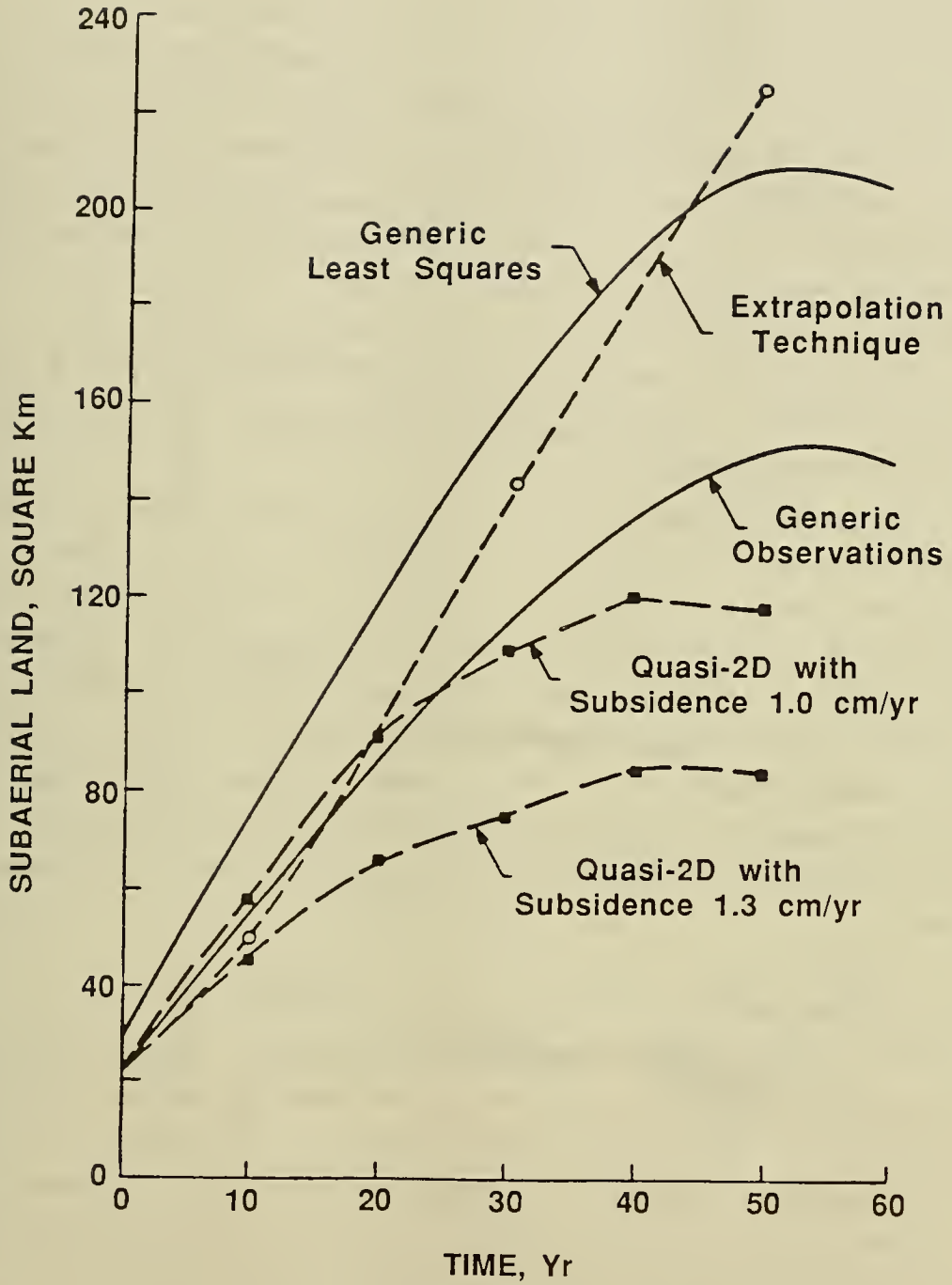

Fig. 10.3. Time Rate of Subaerial Land Growth in Atchafalaya Bay, Louisiana, Calculated by Different Approaches (after McAnally et al., 1984).

Atchafalaya delta land area. The quasi-two dimensional approach employed a one-dimensional numerical model with two subsidence rates to establish a range of possible results and to define sensitivity to subsidence.

Prediction of bed movement due to sediment erosion-deposition was carried out by the application of the two-dimensional, depth-averaged numerical model in Savannah Harbor, Georgia (Ariathurai et al., 1977). The reach of the estuary under investigation was 6,400 m in length, between stations 1 and 3 (Fig. 10.4). The turning basin, near station 2, was the region of heaviest shoaling. Flow and sedimentary data were collected at the three stations previously. Given the flow field (from measurements), the model was verified against measured suspension concentrations. Fig. 10.4 shows predicted bottom evolution with tide for the period 2300 hr on September 24, 1968, to 0500 hr on the following day. The semi-diurnal spring range of tide during this period was 2.6 m. The initial condition corresponds to the situation at 2300 hr when the ebb current was decreasing but there was no measurable deposition. The bottom topography shown in the figure at that time may therefore be considered to represent the bed which was not scoured by the prevailing current. As observed, heavy deposition occurred by 0500 hr, at slack water following flood, particularly in the region of the turning basin. Measured deposit thickness on the order of 1 m there agreed with the model result. Modeling effort such as this one can be used to generate predictive scenarios for sedimentation patterns provided the expected hydrodynamic and sedimentary boundary conditions are well known.

10.4 WETLAND RESPONSE

Shallow bays surrounded by extensive wetlands will expand rapidly in response to a rise both because of the gentle slope and the deterioration of the marshes in response to salinity increases. For example, Barataria Bay, Louisiana, has increased its surface area about 10 to 15 percent over the last century in response to about 1 m of local relative sea level rise (National Research Council, 1987b). In general, however, although wetlands are critically important as a buffer against shoreline erosion, their response to sea level change is complex and not yet fully understood in the quantitative sense.

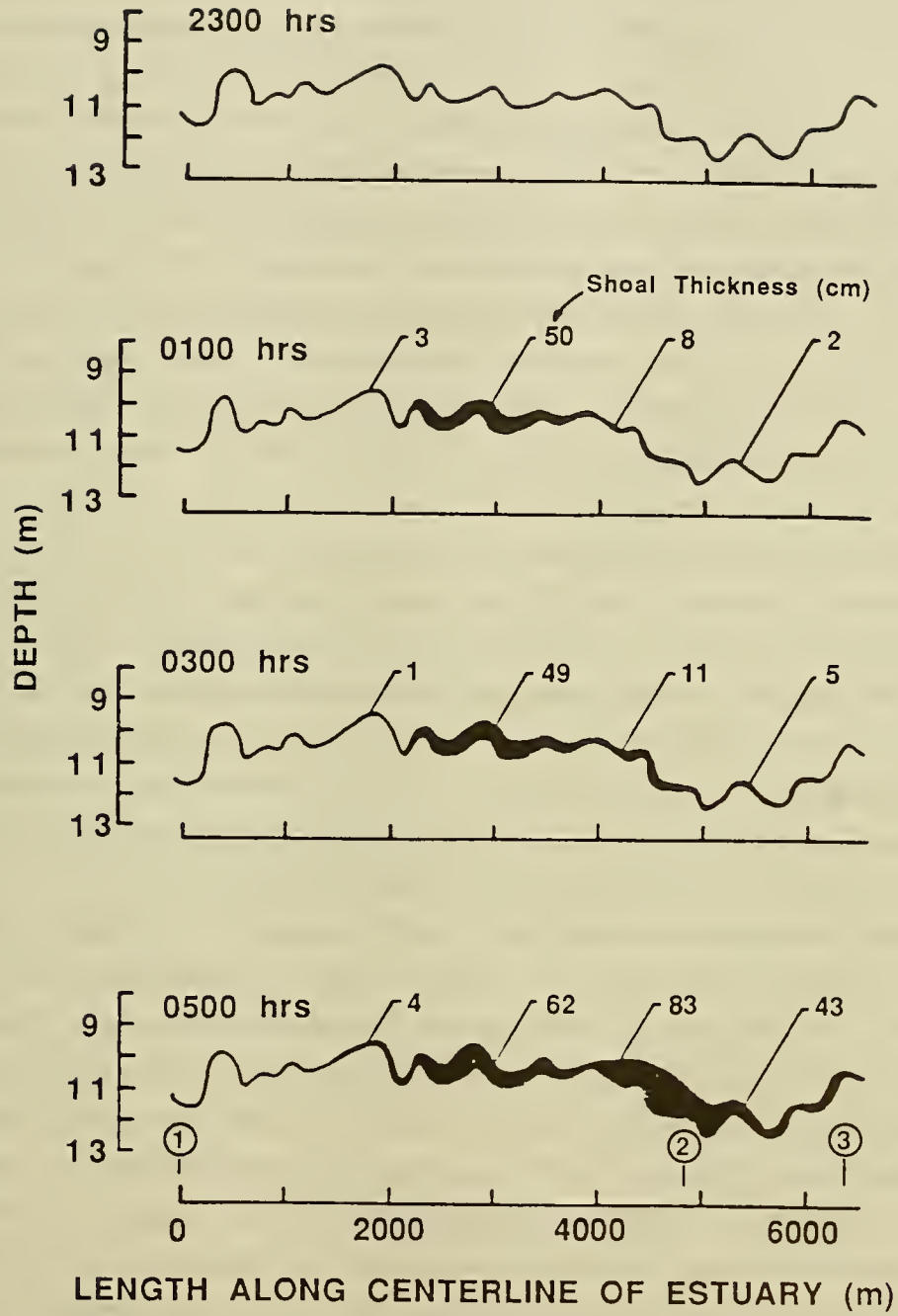

Fig. 10.4. Time History of Bottom Sediment Movement in Savannah Harbor Estuary, Georgia (after Ariathurai et al., 1977).

Wetlands account for most of the land less than 1 m above sea level. These extensive marshes, swamps, and mangrove forests fringe most of the U.S. coastline, particularly along the Atlantic and Gulf coasts. Their estimated original extent in the United States was about 2.0×10^4 km^2 (Hoese, 1967). This area has been significantly reduced through a variety of actions, including an early widespread practice of filling marshlands in urban areas. Wetlands loss has also been caused by other human actions, such as the construction of canals and waterways and the diversion of fluvial sediment to the offshore (National Research Council, 1987b).

Ecological conditions in coastal marshes range from marine to nearly terrestrial. A change in controlling factors, such as water salinity or tidal and wave energy, will cause a displacement in marsh zonation. Generally, coastal marshes are divided into low and high marsh based on their elevation relative to sea level (Redfield, 1972). Since marsh plants are attuned to particular mean water levels, a rise in sea level will shift the distribution of plant species proportionally landward. Beyond this response to variation in relative sea level, however, a more complex set of responses may occur, tied to the type of marsh considered. Thus, anticipated changes in coastal marshes must be assessed within the context of the basic marsh types that characterize the coasts. With respect to the future effects of a rise in sea level, coastal marshes may be broadly divided into backbarrier marshes, estuarine (brackish) marshes, and tidal freshwater marshes (National Research Council, 1987b).

Backbarrier marshes occur along the bayward sides of barrier systems of the Atlantic and Gulf coasts. Studies (e.g., Zaremba and Leatherman, 1986) show that these marshes are formed and destroyed rapidly in such dynamic environments. Maintenance of these marshes therefore appears to be more a function of barrier stability than of the pace of upward growth of the marsh surface, since sediment supplies are ample (Letzch and Frey, 1980). For barriers rapidly migrating landward, there may be a net decline in backbarrier marshes. This has been found to be the case at Assateague Island, Maryland, where sediment blockage by jetties has greatly increased the rate of landward barrier migration (Leatherman, 1983), and the same qualitative result could be anticipated as a result of accelerated sea level rise (National Research Council, 1987b).

Estuarine marshes embrace a wide variety of vegetative species in diverse geologic settings where salinities are less than approximately 30 ppt. These marshes, comprising integral components of major estuarine systems such as the Chesapeake Bay, occur in areas of quiescent waters and ample sediment supply. Accretionary budgets differ widely, as seen from Table 10.1.

The data of Table 10.1 are plotted in Fig. 10.5 in terms of mean marsh accretion rate against relative sea level rise rate, both in mm/yr. It is noteworthy that with the exception of the three data points from Louisiana and one in Georgia (Savannah R.) the remaining data points are confined within a relatively narrow domain bounded by a range of 0.9 to 3.9 mm/yr for sea level rise and 2.5 to 6.1 mm/yr for marsh accretion. The corresponding mean values are 2.3 mm/yr and 4.3 mm/yr, respectively. Thus overall marshes seem to have accreted at almost twice the rate of sea level rise.

The apparent discrepancy between sea level rise rate and accretion rate is most likely to be due to the effect of compaction. The rate of accretion can be thought of as the initial rate of sediment deposition, which depends on the ambient suspended sediment concentration and the sediment settling velocity. In the absence of compaction, and noting that usually in the long run marsh level can at most keep pace with sea level but not rise faster, the maximum marsh accretion rate must equal sea level rise rate. This is indicated by the 45° dashed line (no compaction) in Fig. 10.5. In general, however, compaction cannot be ignored. Laboratory tests on the deposition of relatively thin fine-grained sediment beds (Dixit, 1982) show that the density of the initial deposit (dry sediment mass per unit volume) tends to increase from about 0.05-0.1 g/cm^3 to between 0.2-0.3 g/cm^3 after a few days of relatively rapid consolidation. Beyond this period, further increase in density is very slow. Although there would be a significant difference between these test results and marsh compaction in the field, it is worthwhile to examine the implications of compaction based on the laboratory evidence, qualitatively. Thus, if we assume a two-fold increase in density, the line shown in Fig. 10.5 (compaction) will result. Several data points tend to corroborate the observed linear trend. It can thus be surmised that compaction effects are significant in marshes.

Table 10.1. Rates of Marsh Accretion and Relative Sea Level Rise (adapted from Stevenson et al., 1986)

Location	Mean tidal range (m)	Relative sea level rise (mm/yr)	Salinity (ppt)	Accretion rate	
				Range (mm/yr)	Mean (mm/yr)
Barnstable, MA	2.9	0.9	20-30	3-8	5.5
Prudence Is., RI	1.1	1.9	28-32	2.8-5.8	4.3
Farm River, CT	1.8	1.9	-[a]	-[a]	5.0
Fresh P., NY	2.0	2.2	26	-[a]	4.3
Flax P., NY	2.0	2.2	26	4.7-6.3	5.5
Lewes, DE	1.3	2.0	25-30	-[a]	4.7(>10)[b]
Nanticoke, MD	0.7	3.2	2-6	4.9-7.2	6.1
Blackwater, MD	0.3	3.9	1-5	1.7-3.6	2.6
North R., NC	0.9	1.9	-[a]	2-4	3.0
North Inlet, SC	1.6	2.2	30-35	1.4-4.5	2.5
Savannah P., GA	3.0	2.5	-[a]	-[a]	11.0
Sapelo Is., GA	2.1	2.5	30-35	3.5	4.0
Barataria, LA[c]	0.5	9.5	<1->15	5.9-14.0	7.2
Fourleague, LA	0.3	8.5	10-20	-[a]	6.6
L. Calcasieu, LA	0.6	9.5	15	6.7-10.2	7.8

[a]Not reported.
[b]Lower value obtained by dating with lead-210, higher with cesium-137.
[c]Values based on fresh, brackish, intertidal and water marshes.

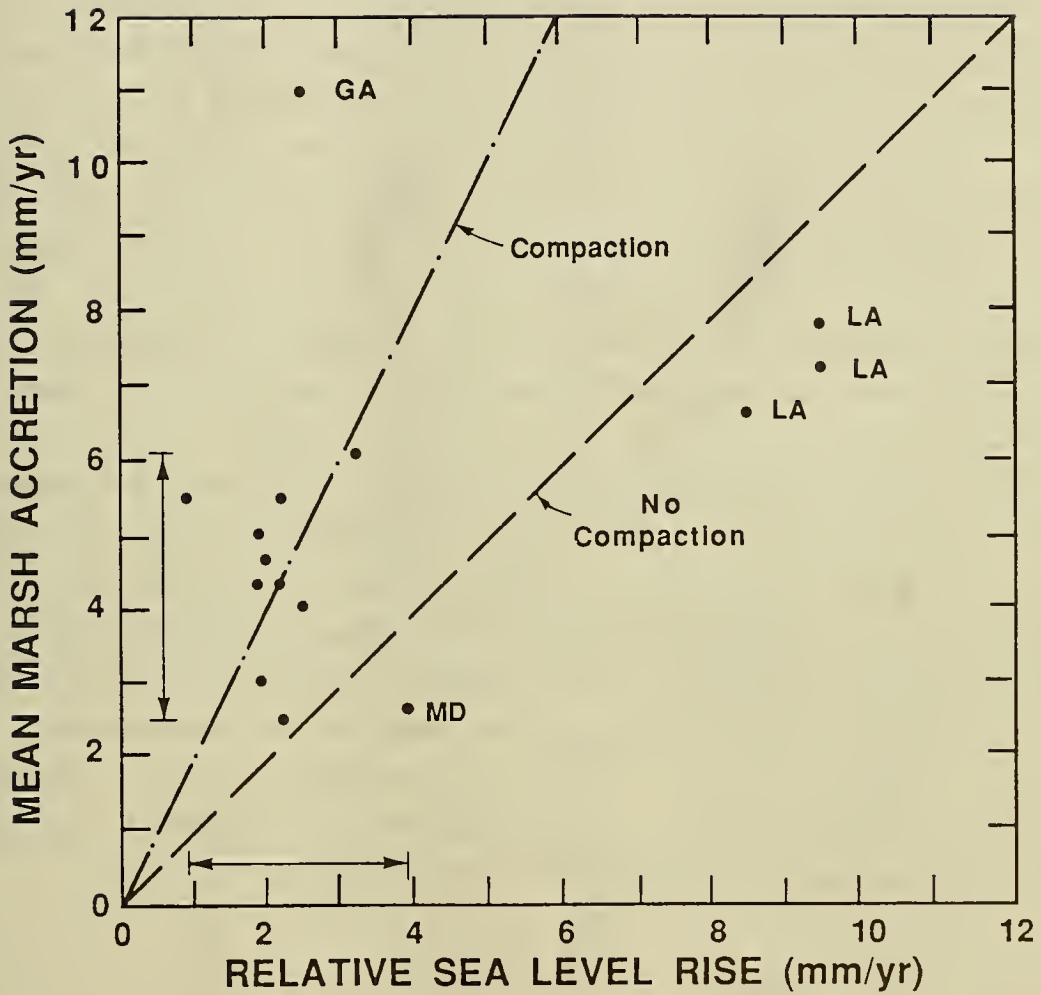

Fig. 10.5. Relationship between Sea Level Rise and Marsh Level Rise Rates (modified, with permission, from Stevenson et al., 1986, "Vertical accretion in marshes with varying rates of sea level rise," pages 241-259 in Estuarine Variability, D. A. Wolfe, ed., copyright by Academic Press, Inc.).

Tidal freshwater marshes are located in the upper reaches of estuaries and other areas where ambient salinities are less than about 5 ppt. The effects of rising sea levels will be saltwater intrusion and the eventual dominance of higher salt-tolerant plants. However, the effects of canalization on tidal freshwater marshes in the Mississippi delta demonstrate that dramatic increases in salinity over a comparatively short period exceed the capability of these marshes to adjust so that rapid losses ensue (National Research Council, 1987b).

The complexity of marsh response to water level and associated factors has meant that predictive modeling effort has been limited. Most modeling to date has been of descriptive nature, as illustrated in Fig. 10.6 (Titus, 1986). Coastal marshes have kept pace with the slow rate of sea level rise that has characterized the last several thousand years. Thus, the area of marsh expanded over time as new lands were inundated. If in the future, sea level rises faster than the ability of the marsh to keep pace, the marsh area will contract. Construction of bulkheads along lagoonal banks may prevent new marsh from forming and result in a total loss of marsh in some areas.

Krone (1985) used a simple deposition model to predict the response of marshes to sea level rise in the San Francisco Bay area. As water leaves a channel and flows onto the marsh surface during a rising tide, the concentration, C, of suspended sediment diminishes as sediment aggregates settle to the marsh surface. As new sediment-laden water from the channel mixes with the previously flooding water, however, the added sediment increases the concentration. This process continues until the tide reaches its maximum elevation after which only deposition occurs while the water drains from the marsh. A mass balance of this process leads to

$$(y_w - y_m)\frac{dC}{dt} + W_s C - C_o \frac{dy_w}{dt} = 0 \qquad (10.4)$$

where y_w and y_m are the elevations of the water surface and marsh surface relative to a selected datum, respectively, W_s is the settling velocity of

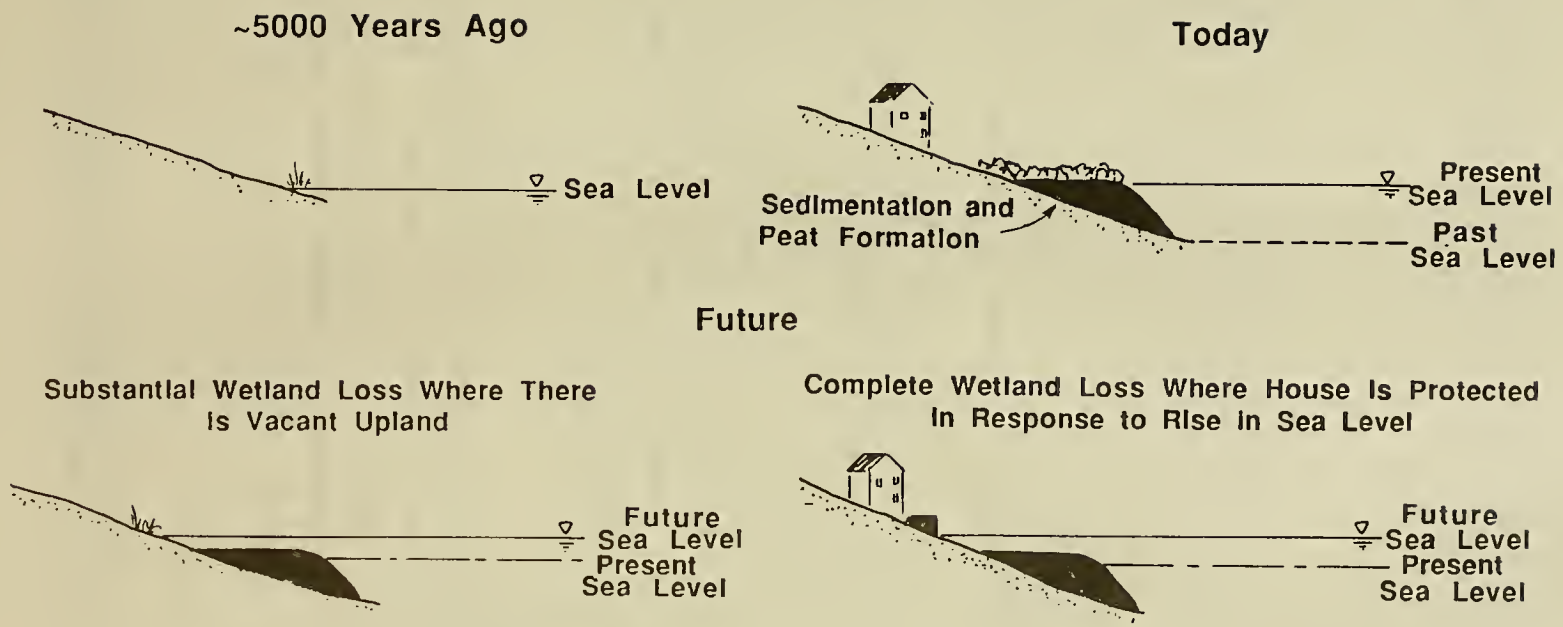

Fig. 10.6. Marsh Evolution with Sea Level Rise (after Titus, 1986).

the suspended aggregates, C_o is the concentration of suspended solids in the flooding waters, and t is time. $C_o = C_o$ during a rising tide, and Co = 0 during a falling tide.

Laboratory tests on San Francisco Bay muds showed that the median settling velocity (by weight) in cm/s of the suspended aggregates is described by

$$W_s = kC^{4/3} \tag{10.5}$$

where k was found to be 110 when the concentration is in g/cm^3. Equations 10.4 and 10.5 are combined to solve for concentration C through a finite difference scheme, and from that the rate of growth of marsh elevation, y_m, given the density of the marsh soil. The value of the ambient concentration in the channel, C_o, was obtained by calibrating the calculated y_m change against measurement of the same at specific sites within the bay.

As would be expected, the rate of rise of marsh elevation is strongly dependent on C_o, as demonstrated by computations for two values of C_o in Fig. 10.7. The computations begin with an initial marsh level (0.15 m) in 1930 when a levee was removed to allow tidal waters to flood the marsh area. The figure shows a more rapid rate of rise during the early period, when the marsh was most frequently flooded, and a slowing rate as a steady rate of rise was approached.

10.5 RESEARCH NEEDS

Present day capability in predicting the evolution of the morphology of the estuarine mouth and adjacent shorelines due to sea level rise is limited. So is our ability to quantitatively evaluate wetland response to sea level rise. Predictive capabilities for sedimentation within the estuary are better, although in areas where high suspension concentrations occur, the physics is poorly understood. Furthermore, the precise nature of chemical and biological variability is not clearly known; hence we are not in a position to establish the quantitative significance of physical versus chemical versus biological control in estuarine sedimentation processes.

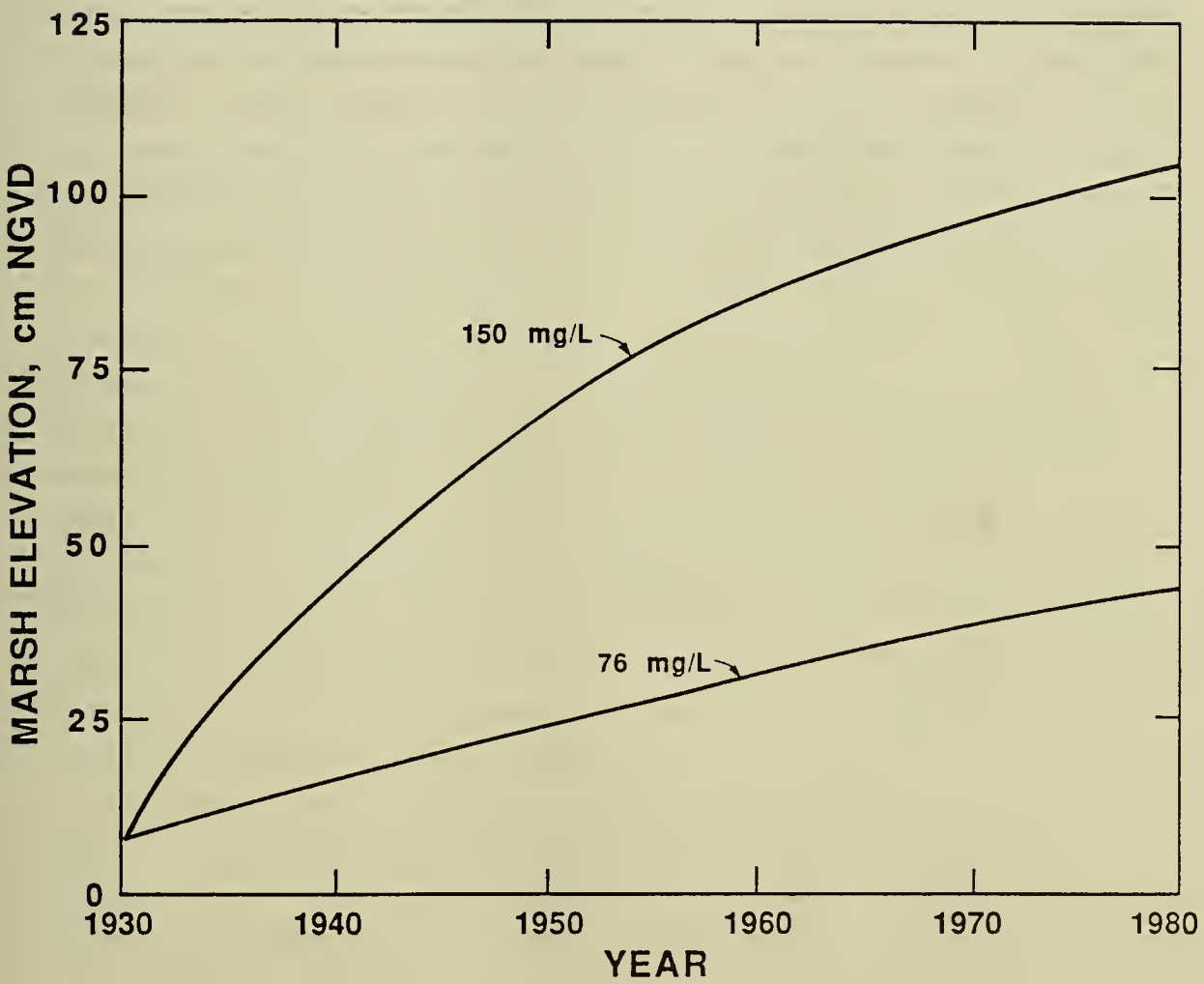

Fig. 10.7. Effect of Suspension Concentration on Marsh Elevation Rise and Sea Level (after Krone, 1985).

As with most other areas, the basic research issues are not particularly related to sea level change, but with the fundamentals of physics, chemistry and biology, as they interact and influence estuarine processes. Advances in knowledge have been diluted by too much site-specific and empirical work. A part of the problem appears to lie with the fact that a great deal of effort has been directed to address specific problems unrelated to basic questions posed by scientists.

In recent years several scientific reports have identified relevant research areas which must be tackled for ultimately improving our predictive capabilities (e.g., North Carolina Sea Grant, "Geophysics in the Environment"; MSRC's "Transport of Fine-grained Sediment"; C. Officer's book; Chesapeake Bay's program; NOAA's estuary program plan).[5] Greater attention needs to be paid to research requirements outlined in these reports. Thus, for example, in a recent report issued by the National Research Council (1987a), research areas in fine sediment transport have been clearly identified.

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[5]Henry Bokuniewicz, Marine Sciences Research Center, SUNY, Stony Brook, New York, personal communication.

11. COASTAL ECOSYSTEMS

11.1 INTRODUCTION

As sea level continues to rise, perhaps at an anthropogenically accelerated rate, existing coastal ecosystems will become submerged and saline water will move inland. Furthermore, tidal amplitude and wave energy may increase due to submergence of protective nearshore reefs and sandbars and other effects (see sections 4 and 5). Water level, water motion, and salinity are principal determinants of the type, nature, and function of coastal ecological systems. Therefore, changes caused by rising sea level could cause dramatic changes in coastal ecosystems.

Coastal ecosystems include open water systems, submerged benthic (bottom) systems, intertidal systems, and supratidal systems. Open water systems consist of plankton (suspended organisms transported by the current), neuston (organisms dwelling in or near the surface film), and nekton (actively swimming organisms). Benthic systems include coral reefs, seagrass ecosystems, and ecosystems of unvegetated sediments. Mud and sandflats that are periodically exposed, and marshes and mangroves are common intertidal systems. Supratidal systems include coastal hardwood hammocks and bottomland hardwood forests. Each of these coastal systems is valued for its contribution of food and cover to the production of a diversity of living coastal resources (Haines, 1979; Peterson, 1981; Zieman, 1982; Odum et al., 1982; Boesch and Turner, 1984; Seaman, 1985). It is likely that a mix of coastal habitats is more important to these resources than any one system alone.

Ecological production (i.e., primary production) in coastal zones is equal to or greater than that obtained with the best mechanized agriculture, yet without the subsidies of plows and chemicals (Odum, 1971). Coastal zones have been said to have natural energy subsidies that together with sunlight account for this very high level of production, namely, the water movement caused by tides, winds, and freshwater discharge (Schelske and Odum, 1961; Odum, 1980).

11.2 ECOSYSTEM RESPONSE

The response of coastal ecosystems to rising sea level can be divided into two components: 1) a vertical growth component, and 2) a horizontal growth component. As sea level rises, an ecosystem capable of growing vertically may stay in the same location. As marine conditions move inland, landward horizontal movement is expected. Under some conditions, seaward expansion of certain coastal systems may occur. For example, intertidal ecosystems that are effective sediment traps, such as salt marshes (Meade, 1982), may not only grow vertically, but may also expand seaward if a large increase in suspended sediment accompanies rising sea level (Frey and Basan, 1985).

Availability of light is a major determinant of the productivity of plankton and submerged coastal ecosystems, such as coral reefs, seagrasses, and mudflats. As sea level rises, open water and submerged benthic communities may not be as productive as they are today in relatively clearwater coastal zones. Increases in tidal range and wave energy may cause an increase in turbidity from suspended sediments. Increased turbidity will reduce the growth rate of light-limited seagrasses (Orth and Moore, 1984; Keesecker, 1986). Intertidal marsh plants, however, may grow better as a result of greater energy subsidies from greater water movement.

The aerial extent of the intertidal zone depends on coastal topography and tidal range. If land rises just inland of the present shoreline, the aerial extent of the intertidal zone will be greatly reduced, unless tidal range increases to compensate. Old shorelines occur just inland of the existing shoreline along much of Florida's coast, for example (MacNeil, 1950).

The organisms that occupy the four general types of coastal habitat (open water, submerged bottom, intertidal zone, and supratidal zone) differ depending on average conditions of salinity, water level, light, temperature, dissolved oxygen, waves, and current (Remane and Schlieper, 1971). Gradients of these factors intersect in coastal zones to give a wide variety of microhabitats. Rising sea level should reposition these gradients (Browder and Moore, 1981).

At any one location, a succession of ecosystems should co-occur with the changes caused by rising sea level. These changes would depend on the

rate of sea level rise and the rate of change of marsh level due to sedimentation or erosion. In intertidal freshwater zones, encroaching salinity will kill bottomland hardwood forests (cypress and gums) and freshwater marsh plants (Odum et al., 1984). These will be replaced by brackish-water grasses and shrubs (Stout, 1984), such as sawgrass (Cladium jamaicense), giant cutgrass (Zizaniopsis miliaceae), giant cordgrass (Spartina cynosuroides), and saltmarsh bullrush (Scirpus robustus). These brackish plants will then be replaced by more saline species of salt marsh plants (Teal, 1986), such as saltmarsh cordgrass (Spartina alterniflora) and black needle rush (Juncus roemerianus). In south Florida, mangroves (Rhizophora mangle, Avicennia germinans, and Laguncularia racemosa) should replace brackish grasses (Odum et al., 1982). Salt marshes and mangroves will be replaced by intertidal mud and sand flats and then by open water over submerged benthic ecosystems (Peterson and Peterson, 1979). If light can penetrate to the bottom and currents are not too erosive, seagrasses will probably develop on submerged sediments.

The extent of inland movement of any particular ecosystem depends on the inland movement of marine conditions and the effectiveness of inland systems to resist invasion or expand seaward. Competition amongst the main structure-defining species of adjacent systems may be important in determining future horizontal extent of each ecosystem.

Island systems and coastal hardwood hammocks will be especially vulnerable to sea level rise.[6] For a circular island or hammock, the area of dry land will be reduced in proportion to the square of the reduction in radius. Low relief islands and hammocks will disappear first along with any habitat value they may have for resident and migratory birds and other wildlife.[7]

Coral reefs may also be particularly vulnerable to sea level rise if the rise is accompanied by an increase in turbidity. In clear water, vertical growth of coral reefs can keep up with rising sea level because of calcification resulting from the photosynthesis of symbiotic zooxanthellae

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[6]T.J. Smith and W.M. Kitchens, comments on Chapter 11, see appendix.

[7]L. Harris, Dept. Wildlife and Range Sciences, Univ. Fla., personal communication.

(Pearse and Muscatine, 1971; Adey, 1978). Turbidity greatly inhibits this process (Loya, 1976). Furthermore, zooxanthellae may leave corals when light, temperature, or other conditions are not suitable for them (Lang, 1971).

To survive, organisms must be able to tolerate or avoid unfavorable conditions. Frequency, amplitude, and regularity (predictability) of fluctuations influence the types of sedentary organisms (e.g., vascular plants, oysters and clams), as well as the ability of vagile organisms (e.g., crabs, fish, and shrimp) to use these habitats. Temporal variation in the coastal environment is most often caused by tides, storms, and seasons. Few species are able to withstand the simultaneous fluctuations in salinity, oxygen, and temperature that occur in many estuaries (Deaton and Greenberg, 1986), though those few that are well-adapted may be able to be very productive because of energy subsidies. Rapid changes in environmental conditions, however, can prevent the full development of an ecosystem. If conditions continually change before any one group of organisms becomes established, then ecological production may be restrained (Montague et al., unpublished manuscript).

Although rising sea level will cause a succession of ecosystems, the rate of rise is probably not rapid enough to prevent full ecosystem development in coastal zones. The seeds and larvae of plants and animals that occur at different points along environmental gradients are widely distributed in estuarine waters. Marshes can become established within one to five years following a sudden appearance of a favorable environment (Montague, unpubl. data). Mangroves may take at most 15 to 25 years to become fully developed (Odum et al., 1982). The rate of sea level rise may or may not be gradual enough to create a significant long-term lag time in the development of successive ecosystems.

Despite the possibility of a timely replacement of coastal ecosystems, whenever a large mass of existing organisms dies, short-term perturbations may accompany the transition to a new ecosystem. Dead plant matter, for example, may temporarily increase in coastal waters and sediment, formerly held in place by roots and rhizomes, may become unstable (Montague, 1986). The added dead matter will decompose, and so may reduce dissolved oxygen levels sufficiently to cause local fish kills. Destabilized sediment will

increase turbidity in coastal waters, which will most likely reduce seagrass and phytoplankton production and may foul the filtering mechanisms of filter feeders such as clams and oysters.

11.3 RESEARCH NEEDS

Prediction of the effects of sea level rise on the future extent of the various coastal ecosystems would be greatly enhanced by the development of a model that incorporates the determinants of the vertical and horizontal growth components of submerged, intertidal, and supratidal coastal ecosystems. Furthermore, to understand the human consequences of changes in extent of each ecosystem, a quantitative understanding of the value of these systems to humans is essential. Energy analysis and natural resource economics techniques (e.g., Odum et al., 1987) may hold promise in this regard if specific fish, wildlife, ecological diversity, and aesthetic values can be included.

Synthesis of existing information on growth determinants and quantative ecosystem values is the first step. The responses of some of the more prevalent coastal plants and animals to some environmental changes are known. As predictions improve the effects of sea level rise on ecologically important physical variables, better predictions of the nature and timing of ecological changes can be made using existing information. For the best predictions, however, new ecological and physiological information may also be required. The most important information needs can be identified with the aid of a literature synthesis and subsequent simulation model. The model will simulate ecological responses to predicted environmental changes (e.g., rising water level, encroaching salinity, increased tidal range and wave energy). Estimates of functions and parameter values will be based on the best available information, but some guesswork is anticipated. Uncertainty in ecological generalizations can be explored by a sensitivity analysis of the model. Needed information can be ranked according to a combination of the uncertainty involved in an estimate and the sensitivity of the model to changes in the estimate. Thus, not only can the most important information needs be identified, but also the consequences of a lack of this knowledge can be demonstrated with the model (Montague et al., 1982).

Although literature synthesis and exploratory models will facilitate the identification of specific research needs, the most relevant research will undoubtedly include several general areas. The time required for full development of subtidal, intertidal, and very nearshore supratidal ecosystems should be established with greater certainty. Under the most rapid sea level rise scenarios, conditions may not remain constant long enough for full development of an ecosystem. If so, the production of fish and shellfish and the stability of shorelines may decline.

Knowledge of nearshore topography and predictions of tidal range are essential to predictions of aerial extent, but so too is an understanding of the level of suspended sediments to be expected and the trapping rate of sediment by subtidal and intertidal plants and microbes (Montague, 1986).

Knowledge of the major regulators of the production of principal animals and plants is essential for coupling predictions of ecological changes to predictions of physical changes. Factors that determine the type and productivities of organisms in the coastal zone include: light (turbidity), temperature, nutrients (including CO_2), salinity, water level, and biochemical oxygen demand (BOD). All of these will be influenced by sea level rise, global warming, and increased levels of atmospheric CO_2.

Physical uprooting and erosion of present ecosystems should be a major agent of ecological change. Predictions are needed both for shores and for tidal creeks. Knowledge of the resistance to erosion of these systems is also required.

Although human values of coastal ecosystems may be compared using various energy and economic analyses, the variation in value within a general category must also be evaluated. Intertidal marshes and mangroves, for example, have been highly touted as good habitat for the growth of juvenile fish and shellfish of commercial and recreational importance. In addition, exchange of materials between the marsh and the estuary is believed to control supplies of nutrients in adjacent estuarine waters. Not all marshes are equivalent in their habitat value, however, and not all exchange significant quantities of materials with surrounding waters (Montague et al., 1987). Perhaps the most important factor in the accessibility of marshes to organisms, and in the exchange of materials, is the density of tidal creeks (Zale et al., 1987). The density of tidal

creeks can be defined as the ratio of length of edge of tidal creeks to surface area of marsh. Knowledge of the influence of creek density on habitat utilization and material exchange may be essential to understanding the relative value of marshes that develop in response to sea level rise. Comparative studies of the effects of creek density have never been reported, however.

Finally, valuable predictions of coastal changes may be obtained from empirical models, if sufficient data can be collected. Four areas of study will contribute. First, paleoecological analysis of cores from various coastal ecosystems can assess responses to past sea level rise (Kurz and Wagner, 1957). Second, analysis of ecological zonation along gradients of salinity and elevation should reflect the kinds of ecosystems to be expected as salinity encroaches and water becomes deeper. Third, analysis of the effects of "experiments of opportunity", in which human or natural events have altered local sea level or caused salinity intrusions, may simulate future effects of sea level rise.[8] Lastly, greater knowledge of the environmental variation under which each major type of system can now exist is needed. Detailed physiometric studies of coastal ecosystems are limited to a few areas, usually near marine research laboratories. Results are often extrapolated to other sites. A given set of predicted environmental conditions, however, may not match those of these few study sites. Each type of coastal system may exist in a much broader range of environments than is now documented, and gradual changes probably occur between system types. Greater regional knowledge of the variety of ecosystem types, and of the variety of environments that support the same ecosystem, will enhance the resolution of empirical ecological predictions.

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[8]T.J. Smith and W.M. Kitchens, comments on Chapter 11, see appendix.

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APPENDIX - WORKSHOP DISCUSSIONS

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GENERAL COMMENTS

This is a well-rounded report; good summary of the subject.

The title should be changed to examine "climate change impacts" rather than "sea level rise impact," to reflect the contents more properly.

Consideration should be given to changes in storm climate resulting from climate change. These impacts could well override the direct impacts of sea-level rise over a 20- to 40-year period. Many examples are available. I am doing research on this topic now (as are others).

Ample consideration must be given to the importance of basic research to address practical problems. For instance, sediment transport theory is in poor shape, with a great need for excellent basic research.

SECT. 2. ESTIMATES OF EUSTATIC SEA LEVEL RISE

There is too little discussion of tectonic contamination of tide-gage records (but I am biased). Otherwise, good points are made regarding noisiness of tide-gage data.

Rather than satellite altimetry, make use of very long baseline interferometry (VLBI) and the differential global positioning system (DGPS) to obtain absolute and relative datums on a global basis. See Carter et al. (1986). Establishment of an absolute network is essential for obtaining a proper separation of sea level vs land level changes.

I see no use for the analysis in Sections 2.2 and 2.3 regarding "noise" in gages. There is no justification for the models used; hence, other than speculation, I see no real use for this arbitrary analysis.

"Simple compaction measurements" advocated in Sections 2.4 and 2.4.2 are not necessarily highest priority. VLBI and DGPS should be considered for this use also.

Increased attention to neo-tectonic contributions to tide-gage results might prove useful. Basic research is required--not just some "weighted averaging," as implied in Section 2.4.1.

New tide gages are required, particularly in the southern hemisphere. However, simultaneous attempts to identify and separate land motions from sea level motions are needed.

We need to use any historical records available to identify the full spectrum of sea level change, so that we can interpret better future sea level measurements. The redness of the spectrum (e.g., Sturges, 1987) argues for adequate long-term measurements.

Modeling must be accomplished where possible. We need improved statistical models to isolate and predict "noise," and dynamic modeling for ice sheet melting/steric expansion to help improve predictions of sea level rise.

SECT. 3. COMPACTION EFFECTS

With reference to measuring compaction (Sect. 3.2), tectonic land level changes and compaction can occur at the same location, so interpretation of compaction data must be done carefully and not be taken to represent the sum of all land level changes.

The discussion of Mississippi River Delta subsidence (Sect. 3.5) is not well balanced.

Bangkok is another excellent example of land subsidence problems. Bangkok is approximately 1 m above sea level; parts are sinking at a rate of 10 cm/yr.

It is not evident that installation of compaction gages is the best use of money. Compaction recorders measure only one element of land level change.

Use VLBI and DGPS for surveying.

SECT. 4. TIDAL RANGE EFFECTS

This section addresses more than tidal range effects. Change the title?

The utility of frictionless estimates is not clear. It is better to use frictional models (even if linearized friction).

Why emphasize resonance effects? What proportion of world coastlines have strong tidal resonance?

The discussion of Mann (1987) in Section 4.4 is misleading and unrealistic. Suggest either complete rewrite or omission of this poor example.

We cannot neglect changes in sediment load and water discharge with climate changes. Climate change may impact these two variables, which in turn will impact tidal range impacts.

Shoaling of estuaries/harbors makes quantification of these tidal range effects difficult. Our knowledge is insufficient to predict large-scale shoaling that can impact tidal range changes.

We need improvement of basic understanding of sediment transport: emphasize basic research and not rely on bulk, empirical sediment transport formulae. The largest potential payback of all coastal engineering might be improvement in understanding the physics of sediment transport.

Tidal range depends not only on growth of parent tides (e.g., M2, S2, N2, K1, O1, S1, etc.) but also on harmonic growth (amplitude and phasing). Our understanding of these nonlinear processes must improve if we are to predict accurately tidal range impacts (de Ronde presented some interesting examples).

SECT. 7. SHORELINE RESPONSE MODELING

This is a key area for setting up responses to sea level rise. Our present understanding is limited to qualitative and semiquantitative models of shoreline response.

The Bruun rule has been the bane of coastal engineers for decades. Geologists have adopted this rule without adequate caution and applied it indiscriminately at times. We need a new and improved understanding of shoreface response to relative sea level rise.

The basic understanding of formation of barrier beaches is qualitative. We do not have a quantitative model for formation of barrier beaches.

The elementary physics of sediment transport, both near bed and suspended load, is inadequate for quantitative models. In a sense, all models now are empirical and qualitative.

Models often are kinematic and not dynamic. Time to reach equilibrium is not considered (time dependence).

Impacts of change in storminess need to be modeled. Perhaps we can use historical data.

Laser technology for profiles does not appear to be the most useful technology.

We need to consider impacts of change in sediment availability and freshwater inflow. How does global climate change impact sediment supply?

The major issue appears to be oversteepening of the inner shelf. This has many people concerned. Is this a valid concern or just a red herring?

Onshore sediment transport is a potential source of stability to beaches. Does onshore (midshelf shoreward) transport really exist? Theory and observations both are ambiguous regarding this factor.

Improved understanding of beach nourishment procedures is required.

SECT. 8. SALTWATER INTRUSION

The impact of varying recharge and discharge from aquifers must be significant, particularly for phreatic aquifers (unconfined) on islands or peninsulas. On Cape Cod, for example, increased discharge may well dominate over any sea level effects. On the other hand, a 1-in. rise in the phreatic surface will cause a depression in the saltwater interface of 40 to 50 in.; increased recharge due to increased precipitation might overcome effects of relative sea level.

Tidal mixing along the saltwater/freshwater interface must be understood better. This mixing is important to salt intrusion as well as biological processes: nutrient exchanges, nutrient budgets, and other biochemical interactions.

Consideration must be given to effects of saltwater intrusion up rivers on groundwater intrusion along the rivers. We cannot address just the coastline.

SECT. 9. UPRIVER SALTWATER PENETRATION

We need increased understanding of the basic physics of intrusion mechanics; e.g., mixing (including small-scale turbulence) across density

interfaces is poorly understood. Bed-form generation and impacts on bed shear stress (hence on intrusion distance) are poorly understood.

Sedimentation within estuaries will impact salinity intrusion. We need better understanding of how sediment supply and transport will vary with global climate change.

SECT. 10. SEDIMENTARY PROCESSES IN THE ESTUARINE REGION

We need to review impacts of global climate change of precipitation, surface runoff, and sediment supply to estuaries.

We must improve our understanding of changes in tidal asymmetries (nonlinear dynamics) in estuaries. These asymmetries will impact not only tidal range, but sedimentation patterns, trapping efficiencies, etc. This is a critical requirement for understanding estuarine sedimentation.

We need improved understanding of near-bed and suspended-load sediment dynamics: basic physics.

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SECT. 3. COMPACTION EFFECTS

With reference to the statement (in Sect. 3.1) "... However, no investigations have been found which identify any specific effects of the inverse problem, i.e., the effects of sea level rise on compaction and subsidence ...," I offer the following addition: Hydroisostasy, the concept that meltwater from Pleistocene ice caps produces loading of the ocean floor that causes significant global deformation, has been examined qualitatively by Daly (1925) and quantitatively by Walcott (1972) and Chappell (1974). The model of hydroisostatic adjustment of continental shelves rests on the assumption that the earth adjusts isostatically to changing surface loads applied over relatively large regions. The deformation through time, in response to changing surface loads, has been refined by Chappell et al. (1982) and has been applied to an area of the Australian coast of North Queensland.

Figure 1 in Chappell et al. (1982) illustrates the effect of a 150-m sea level rise on the deformation of a continental shelf and the adjacent shoreline. As the shallow shelf (roughly 200 km wide) is loaded with 50 m of water, it sinks under the weight of water while adjacent land areas rise (designated by F in Fig. 1a). The effect of the water loading, illustrated in Fig. 1b, is to depress the area seaward of the present coast to a maximum depth of 4 m greater than the eustatic rise.

SECT. 9. UPRIVER SALTWATER PENETRATION

You might want to add an example in Sec. 9.4 from Delaware Bay: Hull and Tortoriello (1979) and Hull (1986) have estimated the salinity intrusion into Delaware Bay resulting from a 13-cm predicted sea level rise for the period 1965-2000. Expressed as the maximum 60-day chloride increase along the Delaware estuary, values as high as 200 mg/L might be expected (Fig. 8 in Hull and Tortoriello, 1979).

SECT. 10. SEDIMENTATION PROCESSES IN THE ESTUARINE REGION

At the end of 10.4 you might want to add ...

Marsh sediments contain more than inorganic matter derived from tidal creeks, though. The vegetation on the marsh surface serves as an autochthonous source of sedimentary material, as most marsh sediments contain 10-90% organic matter. It is not clear how the productivity of these plants may respond to sea level increases or the degree to which transport of organic material from the marsh to the open estuary might be modified by such increases.

Salt marsh plants contain an appreciable quantity of inorganic material. Table A.1 illustrates the composition of some marsh grasses apportioned into aboveground (leaves and stems) and belowground (roots) components.

Table A.1. Annual Production ($g/m^2/yr$) of Delaware Marsh Vegetation (Roman and Daiber, 1984)

	Dry Wt.	Ash	C	N
Aboveground	700	105	273	7
Belowground	5600	1120	2184	56
Total	6300	1225	2457	63

For Delaware marshes, the annual production of ~6000 $g/m^2/yr$ is partially decomposed, partially transported into the open estuary, and partially incorporated into marsh sediments. In Table A.2 we calculate the amount of marsh grasses that must be incorporated into the marsh sediment to maintain the marsh surface against various rising sea levels without any external inorganic sediment supply, given various grass preservation rates. Such a sediment would probably be characterized as a peat but would contain about

20% inorganic ash. If Delaware marshes could retain 60% of their total production, they could be maintained in the face of sea level rises as great as 0.7 cm/yr.

Table A.2. Required Sedimentation ($g/m^2/yr$) by Marsh Grass Alone To Maintain Surface for Various Sea Level Rises

Local sea level rise (cm/yr)	Grass preserved (%)		
	100	80	60
0.3	1500	1800	2100
0.5	2500	3000	4200
0.7	3500	4200	6000
0.9	4500	5400	7600
1.1	5500	6600	9200
1.3	6500	7800	10900

You might also want to add, either to Sect. 10 or to Sect. 11, as appropriate:

Often, protected estuarine shorelines consist of marsh without barrier beaches. These marshy shorelines frequently exhibit high rates of erosion. We often assume, as the shoreline retreats under the influence of rising sea level, that new marsh is created at the landward edge of the transgressing surface. Phillips (1986) has studied the relation between erosion rate, marsh surface accretion, and the slope of the surface being transgressed. He has found that the determination of whether the marsh <u>area</u> will be maintained depends critically on the slope of the surface being transgressed (Fig. 1 in Phillips, 1986). Clearly, the upland slope must be very low (0.2°) and the rate of marsh retreat must be small (0.3 m/yr) to maintain marsh areas, because accretion rates seldom exceed 10 mm/yr (Table 10.1).

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GENERAL COMMENTS

The authors have made a thorough review of the effects of sea level rise on hydrodynamic factors, such as storm surges and waves, and on natural coastlines and man-made structures. I feel that their analysis is complete and thorough, and I agree with their conclusions. More specific comments follow for each section.

SECT. 5. STORM SURGE AND WIND-WAVE RESPONSE

Relative sea level rise affects the heights of storm surges, the height of wind-generated waves, and the reduction of wave height due to bottom friction. An illustration of the magnitude of the effect is shown for each.

Storm Surge

For the case of storm surge, the change in total water level at the shoreline for a constant-depth shelf is due to two causes: the increase in mean water depth, which reduces the hydrostatic response of the shelf due to the applied stress; and the increase in shelf width due to inundation and shoreline recession. The first of these effects can be examined by the governing equation for surge at constant water depth, h_o.

$$\frac{d\eta}{dx} = \frac{B}{(h_o + \eta)} \tag{A.1}$$

Clearly if the water depth, h_o, increases, then the water surface response slope decreases, leading to the obvious fact that the storm surges on very shallow shelves, such as the Gulf of Mexico, are much greater than those on deeper shelves.

It is perhaps more informative to determine relative changes in variables, such as storm surge, due to sea level rise rather than the

absolute change. Using the simple equation for the surge on a constant-depth shelf, Eq. (5.2), in the Mehta et al. report to which this discussion refers, the relative changes can be shown as

$$\frac{d\eta}{\eta} = \left(\frac{-h_o}{\eta+h_o} \quad \frac{dh_o}{h_o} \right) + \left(\frac{B\ell}{(\eta+h_o)\eta} \quad \frac{d\ell}{\ell} \right) \tag{A.2}$$

which shows that the relative change in total water level at the shoreline, $d(\eta + h_o)/(\eta + h_o)$, is roughly equal to the relative change in mean water depth, dh_o/h_o; while the influence of relative change in shelf length, $d\ell/\ell$, is much smaller, since $d\ell/\ell$ is small in itself and is multiplied by a small number. Therefore the surge is most affected by the change in depth.

For the sloping-shelf case, using Eq. (5.7) in the Mehta et al. report discussed herein for the surge height at the shoreline, the relative change in shoreline surge, $d\eta/\eta$, is

$$\frac{d\eta}{\eta} = \frac{\eta-r}{\eta} \quad \frac{r}{\eta} \quad \frac{h_o}{h_o-r} \quad \frac{dh_o}{h_o} \tag{A.3}$$

where r is the dimensional ratio, B/m. For r small (note that for the example with 150-km shelf and h_o = 10 m, r = 0.666), the relative change $d\eta/\eta$ is much less than the change dh_o/h_o, as

$$\frac{d\eta}{\eta} \approx \frac{r}{\eta} \quad \frac{dh_o}{h_o} \tag{A.4}$$

Note that with the assumption of a constantly sloping shelf depth, the increase in shelf width due to the increase in water depth is implicitly assumed. In fact, the difference between the surge height before the sea level rise and the height after the rise can be seen conceptually as simply extending the continental shelf in the first problem farther out to sea until the new water depth due to the sea level rise is met. This extends the shelf at the offshore end of the shelf, where there is little hydrostatic response, so the total effect of the increase in water level is small.

Real continental shelves are likely to be a combination of both of these cases; and it is likely that the response of surge to sea level rise is directly related to the relative change in water depth, indicating that shallow shelves will be much more affected by sea level rise.

Wave Heights

The influences of water depth on wave generation and bottom dissipation are examined in the remainder of Sect. 5. For wave generation on a sloping continental shelf, it is likely that the influence of water depth will be very small, as fetch length for the wave generation will be unchanged. Waves generated locally will not feel the presence of the bottom at the offshore end of the shelf, and thus the generation problem is simply displaced landward by the amount of shoreline recession for a sloping continental shelf.

The influence of water depth on wave dissipation will also not be different for a sloping shelf, assuming that the waves are in deep water at the offshore end of the continental shelf, since again the problem is simply displaced landward and upward.

If, on the other hand, the shelf length is fixed, then the wave height change with water depth can be significant, as shown by the example in the text. This is a very important effect, as stabilized shorelines will have fixed shelf lengths.

Research Needs

In addition to the recommendations in Sect. 5, the most important research problem is to characterize the statistics of hurricane occurrence and landfall locations. If, as some hypothesize, the likelihood of hurricanes will increase, then the present situation needs to be fully documented. Some of this work has already been done in conjunction with the Federal Emergency Management Agency-Federal Insurance Administration studies for coastal counties.

SECT. 6. INTERACTION WITH NATURAL FEATURES AND CONSTRUCTED WORKS

The response of natural and man-made coastal structures to sea level rise is the subject of this section. Natural features such as barrier

islands, shoals, and tidal inlets will be affected greatly, in large part by inundation due to the sea level rise. Natural tidal inlets will be affected by sea level rise principally by the increase in tidal prism due to the increased planform area. If the planform of the back bay remained the same (a bay with vertical sides, which is not very common) and the inlet throat were in equilibrium with the littoral drift on the shoreline, then the principal effect of sea level rise would be the increased shoal volumes needed to keep up with sea level rise. Therefore, there would be increased shoreline erosion, but the hydraulic efficiency of the inlet would remain the same. The deposition on the shoals would occur over the tops of the shoals, which would not produce an areal increase in size.

For most bays, the increase in water level will increase the bay planform, by inundation and by shoreline recession due to increased wind-wave activity in the bay. Therefore the inlet will increase in size. Jetties that line such inlets will be subjected to not only an increased risk of overtopping due the increased water level, but also an increased scour at the toe of the structure or at its head. Indian River Inlet, Delaware, is an example of a jettied inlet that is in jeopardy of losing the jetties as a result of bottom scour due to an increased tidal prism.

For coastal shore and harbor protection structures, sea level rise may be accounted for by designing the structure initially to be modified in the future to account for the intervening sea level rise. This recommendation was made in the National Research Council (1987) report.

It is probably worth noting that there are a few positive results of sea level rise. For example, harbor and maintenance dredging should decrease. Circulation in bay and canal systems should also improve, due to the increased tidal prism.

Research Needs

Clearly, as stated in the text, we need to develop an adequate coastal model, which will allow the long-term prediction of shoreline response to the effects of waves, nearshore currents, and sea level rise. This entails the improvement of our knowledge of nearshore hydrodynamics, nearshore wave mechanics, and nearshore sediment transport. I strongly

support this recommendation, as such a model is needed, even in the absence of sea level rise.

SECT. 7. SHORELINE RESPONSE MODELING

The primary purpose of this section is to discuss the relationship of shoreline recession to sea level rise. The well-known Bruun rule and its progeny are discussed, as are their shortcomings, such as the requirement of offshore sediment transport.

Of utmost importance is the determination of the amount of shoreline recession due to sea level rise. Part of the recession is due to the on/offshore transport of shoreline material. The section points out the importance of alongshore transport as well, with the example of tidal inlets in starving downdrift beaches, leading to shoreline recession, which would occur in the absence of sea level rise.

Clearly we need to document the influence of cross-shore transport under rising sea level conditions. Dean (1987) has indicated that there are problems with the Bruun model. If offshore sources do exist, they may mitigate some of the Bruun rule shoreline recession.

Research Needs

Determine the cross-shore sediment transport paths and the shoreline recession due to sea level rise. Note that this is an integral part of the Sect. 6 recommendation for the development of a general coastal response model.

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SECT. 2. ESTIMATES OF EUSTATIC SEA LEVEL RISE

Previous estimates of the global eustatic sea level change from tide-gage records over the last 100 years cluster between 1.0 and 1.5 mm/yr (Table A.3). Some studies used single, widely-spaced, representative stations (e.g., Fairbridge and Krebs, 1962; Barnett, 1983), while others grouped many stations into a limited number of geographic regions. Global trends were obtained by averaging of regional averages (Gornitz et al., 1982) or by eigenanalysis (Barnett, 1984; Aubrey and Emery, 1983). However, the latter technique applied to sea level data may have some drawbacks (Solow, 1987).

More recently, Gornitz and Lebedeff (1987) derived a global sea level trend using a similar averaging approach, outlined in Sect. 2.3 of the Mehta et al. (1987) report discussed herein. Vertical movements determined from ^{14}C-dated Holocene sea level indicators (Pardi and Newman, 1987) were removed (see below). The global mean corrected sea level change is 1.0 ± 0.1 mm/yr (95% confidence interval). The arithmetic mean of corrected least-squares slopes on individual stations is 1.2 ± 0.3 mm/yr (Gornitz and Lebedeff, 1987). Sea level is rising in all but three regions that are characterized by small station populations or sparse long-range data (Fig. A.1). Nevertheless, large regional variations in sea level remain, even after subtraction of long-term trends. Some of the sources of noise in mareograph data (Table A.4) and means of removing them are briefly discussed below.

Data Quality and Distribution

Tide-gage records often are too short, are variable in length, contain data gaps, and are geographically biased. Only around 300 tide stations from the Permanent Service for Mean Sea Level (PSMSL) have usable records of 20 years or longer. Although nearly 90% of the stations lie north of 23°N, this represents over 70% of the world's total coast length.

Table A.3. Estimates of Sea Level Rise from Various Sources

Rate (mm/yr) Uncorrected	Corrected	Comments	References
>0.5		Cryologic estimate	Thoraninsson (1940)[a]
1.1 ± 0.8		Many stations, 1807-1939	Gutenberg (1941)[a]
1.2-1.4		Combined methods	Kuenen (1950)[a]
1.1 ± 0.4		Six stations, 1807-1943	Lisitzin (1958)[a]
1.2		Cryologic estimate	Wexler (1961)[a]
1.2		Selected stations, 1900-1950	Fairbridge and Krebs (1962)
3.0		Many stations, 1935-1975	Emery (1980)
1.2		Many stations, grouped into regions, 1880-1980	Gornitz et al. (1982)
	1.0	Many stations, grouped into regions	Gornitz et al. (1982)
1.5 ± 0.15		Selected station, 1903-1969	Barnett (1983)
1.4 ± 0.14		Many stations, grouped into regions, 1881-1980	Barnett (1984)
2.3 ± 0.23		Many stations, grouped into regions, 1930-1980	Barnett (1984)
0-3		Not specified	Aubrey (1985)
	1.0 ± 0.1	Mean of regional means, 1880-1982	Gornitz and Lebedeff (1987)
	1.2 ± 0.3	Arithmetic mean, 1880-1982	Gornitz and Lebedeff (1987)
	1.1	East Coast only	Peltier (1986)

[a]In Lisitzin 1974, after Barnett 1983.

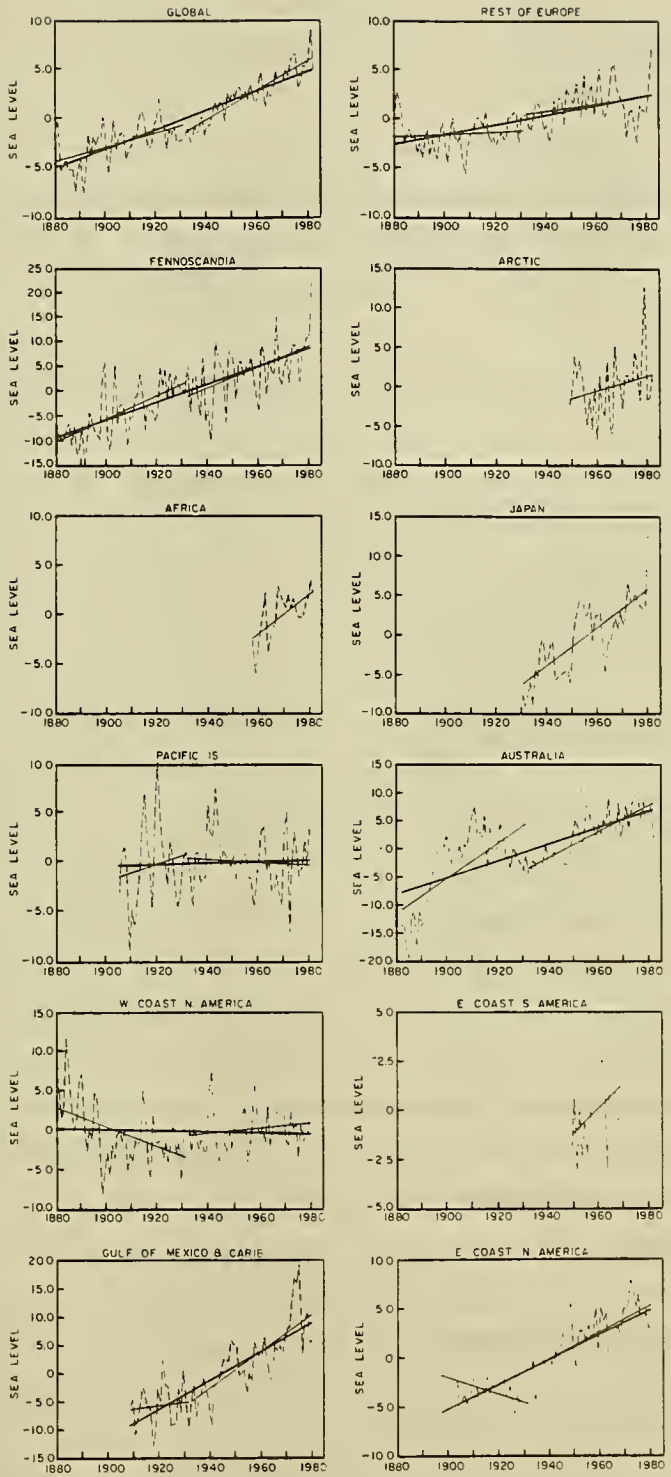

Fig. A.1. Global and Regional Sea Level Curves, 1880-1980. Heavy line represents least-squares slope for the past century. Lighter lines represent slopes for the intervals 1880-1931 and 1932-1980. Units are in centimeters (from Gornitz and Lebedeff, 1987).

Table A.4. Summary of Processes Affecting Sea-Level Changes

Process	Rate (mm/yr)	Amplitude (cm)	Period (yr)
Glacioeustasy	Up to 10		Around 7000 following deglaciation
	~1		Last 100
Vertical land movements			
Long-wavelength processes, 100-1000 km			
Glacioisostatic changes	±1-10		10^4
Shelf subsidence due to oceanic lithosphere cooling and sediment/water loading	0.03		10^7 to 10^8
Shelf sediment accumulation, global	0.02-0.07		10 to 10^6
Short-wavelength processes, <100 km			
Neotectonic uplift/subsidence	±1-5		10^2 to 10^4
Shelf sediment accumulation, local near large river deltas	1-5		10 to 10^4
Anthropogenic activity			
Water impoundment in dams and reservoirs	-0.75		<100
Groundwater mining (to river runoff)	≤0.7 (?)		<100
Subsidence due to groundwater/oil/gas withdrawal, local.	3-5		<100
Oceanographic and atmospheric effects			
Geostrophic currents		1-100	1-10
Low-frequency atmospheric forcing		1-4	1-10
El Niño		10-50	1-3

Geographic bias can be further reduced by grouping stations into a limited number of regions, weighting each region by the area of ocean covered, the shore length, or the relative data reliability. Existing tide stations should be maintained in order to extend the record length. New sites should be located in stable areas that are presently underrepresented. Methods of improving the tide-gage network and linking to satellite altimetry will be described in the discussion of Sect. 3.

Vertical land motions (glacioisotasy and hydroisostasy, neotectonism, sediment loading)

Glacioisostatic movements have been removed from U.S. tide records by applying viscoelastic earth models (Peltier, 1986). An alternative approach is to use late Holocene (≤6000 yr) sea level curves derived from ^{14}C-dated sea level indicators (Pardi and Newman, 1987). These curves include glacioisostatic and local neotectonic movements. Therefore, subtraction of late Holocene sea level trends from tide-gage trends should minimize these effects (Gornitz et al., 1982; Gornitz and Lebedeff, 1987). The two approaches are not completely independent, in that Peltier uses ^{14}C data to calibrate his models.

Sediment loading and lithospheric cooling (Pirazzoli, 1986) are insignificant causes of global sea level change on a 100-year time scale (Table A.4). The subsidence rate along the U.S. East Coast continental shelf and margin over the last 12 million years is only around 0.03 mm/yr (Heller et al., 1982; Steckler and Watts, 1978). The global sea level rise corresponding to present deposition of river sediment on the continental shelves and slopes is only 0.01-0.02 mm/yr (after Milliman and Meade, 1983; Holeman, 1968). These rates are only a fraction of the total observed sea level rise. However, locally, near major deltas (such as the Mississippi delta), relative sea level curves may show much higher rates (e.g. ~9 mm/yr along the Louisiana coast).

Anthropogenic activity (dam building, enhanced sedimentation, subsurface fluid withdrawal)

Impoundment of water behind dams and reservoirs and infiltration of water into aquifers, between 1932 and 1982, are equivalent to withholding a

potential sea level rise of ~0.75 mm/yr (Newman and Fairbridge, 1986). Ground-water mining, which increases runoff to the oceans and thereby tends to raise sea level, could counteract a large fraction of this effect (Korzun, 1978).

Subsurface fluid withdrawal (oil, gas, water) can locally exacerbate the eustatic sea level rise. For example, subsidence rates of around 4.2 mm/yr have been measured between 1906 and 1978, at the Galveston Island tide gage (Gabrysch, 1984). Here, compaction due to groundwater removal could account for as much as 2/3 of the relative sea level rise. On the other hand, at Savannah, Georgia, where land subsidence of up to 4 mm/yr has been reported inland between 1955 and 1975, benchmarks near the tide gage at Fort Pulaski (on the shore) show no subsidence during this period (Davis et al., 1977). These subsidence effects are thus highly variable and site-specific. This may necessitate the establishment of the local subsidence history for each tide station. Independent means of estimating land subsidence will be reviewed in the discussion of Sect. 3.

Atmospheric and oceanographic effects (winds, waves, currents, ocean temperatures, and salinity)

Atmospheric and oceanographic processes generate considerable interannual sea level variability. However, monthly and annual averages of sea level records greatly reduce the <1-year sources of variability (due to tides, storms and seasonality). Longer fluctuations, up to a decade in duration and coherent over long distances, are produced by steric changes (temperature or salinity), ocean currents, and the El Niño-Southern Oscillation.

Multivariate regression can be used to minimize large-scale, low-frequency atmospheric forcing (winds, currents). Large-scale spatially coherent signals (such as those associated with the El Niño phenomenon) can be identified and quantified using eigenanalysis (Chelton and Enfield, 1986).

SECT. 3. COMPACTION EFFECTS

As noted above, compaction due to ground fluid withdrawal can affect relative sea level curves. This impact should be greater for cities

located on unconsolidated coastal plain sediments than for cities on incompressible crystalline bedrock. This hypothesis can be tested for the North American Atlantic Coast. The mean sea level trend for 24 coastal plains stations between Key West, Florida, Long Island, New York, and Cape Cod is 2.90 ± 0.74 mm/yr, as compared with 2.46 ± 0.59 mm/yr for 16 stations on crystalline bedrock between Connecticut and St. John's, Newfoundland. However, the difference between the means is not significant at the 95% confidence level, using Students' t-test. Removal of the long-range trends does not alter this finding. Therefore, compaction effects may not be the most important factor in the observed subsidence of the East Coast.

Whatever the ultimate cause of subsidence, it is desirable to isolate vertical land motions from the eustatic sea level signal. Absolute sea level positions must be measured relative to an ultraprecise geodetic reference system, such as very long baseline interferometry (VLBI) and the global positioning system (GPS) (Carter et al., 1986). VLBI will use a network of fixed and mobile radio telescopes for precise geodetic measurements, accurate to the 1-cm level, by the 1990s. The reference point will be the earth's center of mass, as determined by satellite laser ranging, over time. The VLBI network will be tied in with the GPS, which employs portable ground-based radio receivers to detect satellite radio transmissions. By using phase differencing, the GPS signal can be applied to geodetic observations, in a manner similar to VLBI. VLBI/GPS will be connected to a global network of around 100 "absolute" sea level stations (GLOSS); and ancillary meteorological and oceanographic variables that affect sea level will also be monitored.

Preliminary comparisons of sea level positions from tide gages and GEOSAT altimetry (Wyrtki, 1987) show a fairly good match, although only covering a small area of the equatorial Pacific Ocean, over a 5-month period. This experiment illustrates the potential of satellites for sea level measurements, even with current technology. This approach can be extended to "calibrate" satellites for longer-term, larger-scale observations over wide areas of oceans that lack tide gages.

SECT. 7. SHORELINE RESPONSE MODELING

The shoreline response to sea level change can be investigated using historical data. Historical U.S. shoreline changes have been compiled into a computerized data base, the coastal erosion information system (CEIS), based upon measurements from old maps, charts, and aerial photographs, covering at least a 40- to 50-year period (May et al., 1982, 1983). The data have been averaged into 3′ x 3′ latitude-longitude grid cells. While spatial coverage is fairly continuous over large areas, the shoreline displacement history typically does not have more than 6 or 7 time measurements. Rates of less than 0.6-1.0 m/yr (over a 25- to 40-year interval) lie within the uncertainty. Other problems with data quality have been encountered. Therefore, it may be desirable to check the CEIS data against independent measurements. Furthermore, the data show high spatial variability. The United States is covered by 71 tide-gage stations with records covering 20 years or longer, which have fairly continuous time series but are spatially discontinuous.

Initially, the records for each tide-gage station can be matched with the corresponding CEIS cell that encloses it. In order to obtain more extensive coverage, sea levels can be averaged over coastal segments demonstrating a similar sea level behavior, or by linear interpolation between stations. The CEIS data can be smoothed further by using 3- to 5-cell running means. The optimum smoothing scale can be assessed by sensitivity studies. The shoreline displacement can be divided into both inundation and erosion components. The relative contribution of inundation or erosion can be assessed at each tide station, or along coastal segments. The observed average displacement rate can then be compared with those predicted by various models, such as the Bruun rule, or a sediment budget analysis, for historical rates of sea level rise.

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SECT. 6. INTERACTION WITH NATURAL FEATURES AND CONSTRUCTED WORKS

Text Discussion

The text correctly notes that "The major ... question ... is whether to include projected long-term estimates or not." Design of many constructed works is based on economic justification of incremental benefits vs incremental costs. Present practice usually includes a straight-line extrapolation of present rates of apparent sea level rise (sea level rise plus subsidence) to account for future changes. The costs of building additional structure freeboard to account for accelerated apparent sea level rise can be high; they are justifiable only when either the projected rate is known with some confidence or if the potential risks of not building it are too great to bear.

Unfortunately, raising structures some time after their construction can be extremely difficult. Berm costs include not only construction costs, but also those to purchase the land for extra base width. The latter can be politically and socially unsatisfactory when the areas adjacent to the structures have become developed. Other measures, such as advance purchase of easements and innovative construction techniques, may be needed.

Remarks about groin flanking problems also apply to transverse training structures in tidal rivers (which usually work better than beach groins). Flanking of these structures is a significant problem that sea level rise may exacerbate, but overtopping of dikes may not be a problem. Sometimes training dikes work better when overtopped.

Roy Beard (1987) has noted that the present established (by law and precedent) practice of designing for specific flood levels (i.e., a design storm) is not consistent with the goal of flood protection for the common good. A more rational approach is design to protect against a specific level of economic or social damage at given risk levels. This applies

equally well to floods resulting from increased sea level. While overcoming precedent to change design criteria for traditional flood events may be impossible, applying these concepts to sea level rise damages may be an opportunity for a different approach.

Drainage structure effectiveness will be compromised by higher sea level. Responding to this effect will require modifications to such structures and perhaps addition of pumping capacity.

Locks are frequently used to limit salinity intrusion in navigable waterways. Rising sea level can endanger the gates of locks, requiring gate replacement and perhaps raising of the lock walls. Limiting freshwater loss and saltwater intrusion through locks is a major design concern that will certainly grow worse with higher sea levels.

As climate changes and precipitation and runoff patterns change, river floods could become worse or better. Hurricanes could become more or less common at a particular site. Thus these changes could affect structure design requirements as much as or more than sea level rise.

Research Needs

Structure design and evaluation procedures can be revised to account for accelerated sea level rise, but they are practical only if reliable and precise estimates of future sea level rise rates are commonly agreed to. Then economically sound decisions can be made. The present broad range of possible rates of rise does not encourage design engineers to begin including any of them; therefore, additional work is needed to refine the estimates and obtain a consensus on them.

Design criteria such as damages related to specific rates of sea level rise and specific probabilities of occurrence are needed. Better mathematical descriptions of the processes by which earth mound structures are overtopped and eroded will be needed if the methods of setting protection criteria are changed.

Geotechnical and structural research is needed to develop ways to more economically raise protection levels.

Designs are needed to reduce flanking of perpendicular structures such as groins and dikes.

The effect of structures on increasing local subsidence rates should be quantified, so that designs can include all components of apparent sea level rise.

Research is needed on methods to limit saltwater exchange through locks, so as to improve their effectiveness.

Much more research is needed to quantify the climatic consequences of the greenhouse effect. This includes changes in rainfall and runoff plus the frequency and strength of tropical and extratropical storms, since these can affect structure design and planning as much as sea level rise does.

SECT. 8. SALTWATER INTRUSION

Text Discussion

Climate changes may have an even more profound effect on saltwater intrusion than sea level rise has. Drier weather in agricultural areas will lead to increased groundwater pumping, and wetter weather will have the reverse effect. Recharge rates will be affected, but with a time lag that may be considerable.

Cost of water can completely change water use patterns; therefore, many projections that seem reasonable now may be inoperative in 25 years, as water becomes more scarce and as saltwater intrusion threatens supplies.

If riverine intrusion (also subject to climate changes) changes water supply or impoundment practices, groundwater use will be affected. If river supplies are abandoned because of saltwater intrusion, there is likely to be increased demand for groundwater sources, exacerbating the groundwater problem.

Tertiary effects may occur, such as occurred in St. Louis. In that case a large decrease in groundwater pumping (caused by movement of industry) let the groundwater level rebound to previous high levels, flooding an area that had become highly developed during the period of low water table. If saltwater contamination leads to abandonment of groundwater pumping over a large area, similar responses could occur in coastal areas.

Research Needs

Simple models can definitely be useful for sensitivity studies of various scenarios, including different sea level stands, recharge rates, and pumping rates.

Knowledge of model reliability is needed, particularly when the subsurface geologic structure and boundary conditions are poorly defined.

Better evaluation of climatic effects is needed, so that various scenarios (including status of river supply of fresh water) can be tested in models.

Models to demonstrate water availability should be combined with econometric models, to evaluate societal responses to changes in water availability and costs. Optimization modeling should be demonstrated for effectiveness in combining all water sources for minimum cost.

Barrier technologies -- injection, extraction, and geotechnical -- can be explored for improved methods and materials.

Alternative water sources should be explored, particularly inter-basin diversion and desalinization, if econometric models show that huge price increases in water may be required to deal with shortages.

SECT. 10. SEDIMENTATION PROCESSES IN ESTUARIES

Text Discussion

Atchafalaya Bay delta growth results, as mentioned in the text, are highly dependent on apparent sea level rise. Recent results with more carefully developed, spatially variable rates confirm the effect. The effect extends to flood protection designs, which are markedly different with and without projected rise.

Plant die-off in marshes and marginal areas, caused in part by saltwater intrusion, will dramatically affect erosion and deposition of shorelines.

Sediment supply rates are changing, which is another primary reason for Louisiana land loss. This effect could be exaggerated by climate changes or by different water resource policies caused by sea level rise and rainfall changes.

Redistribution of sediments and generation of new sediment sources and sinks will occur, altering dredging requirements by changing shoal

locations. If the principal locations of dredging are moved inland, dredged material disposal could become even more costly. These circumstances could conceivably lead to new solutions that minimize the total cost of shore protection, dredging, navigation, and water supply, such as reducing channel depths by use of lightering and creation of artificial shorelines and port islands with dredged material.

Research Needs

Both site-specific research and generic research are needed to work both ends of the sedimentation question: the lack of sediment that contributes to shoreline loss and the excess of sediment that impedes navigation.

For sedimentation purposes, exact rates of sea level rise are not necessary. Approximate values will do for impacts evaluation. Local subsidence rates, which are not very well defined, are important and should be better defined. Better datum definition and water level measurements are needed.

Evaluation of sea level rise impacts should also address changes in water and sediment supply, which may be changing as dramatically as sea level.

Demonstrations are needed of total cost minimization for all the consequences mentioned in the last paragraph in the text discussion above. This will involve modeling sediment, saltwater intrusion in rivers and groundwater, shoreline response, and shipping.

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SECT. 8. REVIEW OF IMPACTS OF SEA-LEVEL RISE ON GROUNDWATER RESOURCES

Introduction

In coastal aquifers, groundwater flows from inland recharge areas to discharge areas at or near the coastline. Due to the presence of seawater in the aquifer formation under the sea bottom, a zone of contact, or interface, is formed between the lighter fresh water and the underlying, heavier seawater. The location of the interface is determined by the physical and hydraulic properties of the aquifer, the amount of recharge and flow in the aquifer, and whether pumping from inland well fields reduces the slope of the piezometric surface and the seaward flow of fresh water. A rising sea level will cause the saltwater-freshwater interface to move inland, increasing the salinity of coastal aquifers and threatening coastal well fields. Measures such as hydraulic or physical barriers may protect coastal aquifers in some cases, but abandonment and moving farther inland may be the only reasonable, economical solution in other cases.

Hydrologic Setting in Florida

Impacts of Previous Higher Stands of the Sea. Pleistocene-age shorelines that occur at elevations higher than present-day sea level serve to indicate that sea levels have been higher during some periods in the past. In Florida, at least four Pleistocene shorelines are recognized (MacNeil, 1950): the Okefenokee [150 ft above mean sea level (MSL)], Wicomico (100 ft above MSL), Pamlico (25-35 ft above MSL), and Silver Bluff (8-10 ft above MSL) shorelines. The Silver Bluff shoreline is correlated with the climatic optimum, a period approximately 9100 years before the present, when the temperature was about 5°F higher than the present and sea level was on the order of 10 ft higher than the present sea level (Brooks, 1970).

The long-term water-quality impacts that the higher stands of the sea have had on the Floridan Aquifer are still evident in a large part of the aquifer, particularly where it is confined and occurs at depth. Salt water occurs in the Floridan Aquifer in south Florida, along the east coast of Florida to just south of St. Augustine and in the St. Johns River valley, and along the west coast northward almost to Tampa Bay (Stringfield, 1966). This water appears to be seawater that has not been flushed from the aquifer. Some of it is connate (formation) water, but some of it entered the aquifer during higher stands of the sea during Pleistocene time. In southwestern Florida, dissolved solids in the aquifer increase downgradient, and the general flow pattern in the aquifer is that of flow toward the south and west that has been gradually improving groundwater quality farther and farther downgradient in the aquifer since the last period of marine inundation (Steinkampf, 1982).

Saltwater-Freshwater Interface Location. The location of the saltwater-freshwater interface in the Floridan Aquifer apparently represents an approximate equilibrium between the existing piezometric heads in the aquifer, aquifer flow rates, and discharges to the sea. In the Floridan Aquifer in northeast Florida, piezometric heads in the aquifer are still above sea level at the coastline and eastward beneath the Atlantic Ocean, and the saltwater-freshwater interface is inferred to extend eastward more than 70 miles beneath the Atlantic Ocean (Brown, 1984). In west-central Florida, the intersection of the interface with the top of the Floridan Aquifer probably occurs offshore, and the intersection of the toe of the interface with the bottom of the aquifer probably ranges from a few miles to about 20 miles inland from the coastline (Wilson and Gerhart, 1982).

In the Miami area, the Biscayne Aquifer is unconfined, and the saltwater-freshwater interface occurs near the coastline, where piezometric heads in the aquifer are at or near sea level. The interface in the Biscayne Aquifer consists of a broad zone of transition that is probably about 2000 ft wide at the base of the aquifer (Cooper, 1964). Saltwater intrusion has occurred inland from the coast and along tidal streams and canals that have lowered groundwater levels by means of excessive drainage and improperly placed control structures (Klein and Hull, 1978).

Analytical and Numerical Models

Analytical and numerical models to describe the location of the saltwater-freshwater interface have been developed for many different applications. Most of the analytical models are based on the Ghyben-Herzberg relation, which predicts that the location of the interface below sea level is approximately 40 times the value of the freshwater head above sea level, and on the Dupuit assumption, which assumes horizontal flow in the aquifer. The location of the interface can be predicted for both confined and unconfined aquifers (Bear, 1979). Two-dimensional solutions based on complex variables and conformal transformations have been developed (e.g., Henry, 1964; Van Der Veer, 1977). The lateral and vertical movements of the interface in response to pumping can be analyzed using solutions developed by Strack (1976) and Schmorak and Mercado (1969), respectively. The effects of pumping from the saltwater side of the interface to increase the allowable pumping rate from the freshwater side of the interface have been investigated by Vandenberg (1975). Numerical solutions have been developed that allow dealing with heterogeneities in aquifers and with two-dimensional and transient conditions (e.g., Shamir and Dagan, 1971; Segol et al., 1975; Segol and Pinder, 1976; Mercer et al., 1980). Three-dimensional effects have also been considered (Huyakorn et al., 1987).

Protective Measures

Measures to protect coastal well fields from saltwater intrusion generally can be classified as physical or hydraulic barriers (Kashef, 1986). Physical barriers include sheet piles, cofferdams, and grout curtains to retard the landward movement of seawater. Hydraulic barriers include recharge wells and spreading basins, which maintain a pressure ridge along the coastline perpendicular to the direction of saltwater intrusion, and discharge or extraction wells, which create a continuous pumping trough with a line of wells adjacent to the sea (Todd, 1980). Case histories in Long Island, California, and Florida (Todd, 1980) illustrate the impacts of saltwater intrusion and describe a number of

mitigative measures intended to protect freshwater resources near the coastline.

Research Needs

The long-term persistence of Pleistocene-age water-quality impacts in parts of the Floridan Aquifer indicates that parts of the aquifer system may respond more slowly than expected to changes in head and water quality due to sea level rise, at least in areas where the aquifer is confined and at great depth. Thus, the transient response of the saltwater-freshwater interface in deep confined aquifers needs to be better understood to determine whether the predicted sea level rise will have any significant impact on such systems. Also, many of the analytical and numerical models in the literature are limited to describing the location of the saltwater-freshwater interface under equilibrium conditions, or they do not describe in detail the response to mitigative measures. Thus, an evaluation of the analytical and numerical models in the literature to determine which existing models can be used to investigate transient conditions and various protective measures is needed. In addition, the large number of case histories and descriptions of protective measures could be analyzed systematically to evaluate the effectiveness of these measures and provide guidelines for site-specific applications in areas where saltwater intrusion is expected to be a significant problem.

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SECT. 10.4 WETLAND RESPONSE and SECT. 11. COASTAL ECOSYSTEMS

Two important coastal ecosystems were excluded from consideration in Sect. 11. These are coral reefs and tropical hardwood hammocks. Sea level rise would affect both of these systems. Some 7000 years before the present, an extensive barrier reef system was located offshore of Fort Lauderdale. With the small rise in sea level since then, this reef system has been extirpated and the present Florida reef track has developed (Lighty et al., 1978; Jaap, 1984). Although tropical hardwood hammocks are usually found inland from the coast (Olmstead et al., 1980) they occur at elevations of 1-5 m above MSL. Thus a rise in sea level would have an influence on these systems as well.

Most conventional wisdom states that as sea level rises, affected coastal ecosystems will simply shift inland. This may not be the case. A man-made alteration of tidal regimes, water height, and salinity gradients along the Savannah River mimics potential sea level rise and provides data that can be used in evaluating ecological responses to sea level rise. Briefly, salt-tolerant marsh grass species (e.g., Spartina cynosuroides, Scirpus robustus) found in the lower portions of the Savannah River quickly invaded affected upstream areas of the river that had been dominated by tidal freshwater species (primarily Zizaniopsis miliaceae). These tidal freshwater species, however, did not shift their own ranges farther upriver. They exist now within their former range but in greatly reduced abundances. Thus the tidal freshwater marsh community was reduced in areal extent within the overall system (Latham et al., 1988). Interesting questions concerning ecological effects of sea level rise emerge: Did the response of the vegetative communities along the Savannah arise because the natural environmental gradients were steepened rather than just shifted

inland or because the mid-river communities (the tidal freshwater marshes) were less resilient than the other community types? Because sea level rise is a disturbance, the nature of each coastal ecosystem's response to disturbance must be understood before predictions can be made regarding its response to sea level rise.

An important influence of sea level rise would be the introduction of sulfur and its biogeochemical cycles into previously freshwater regions. This is of particular relevance to microbial decomposition of organic matter. Methanogenesis is a dominant pathway in freshwater and sulfate reduction in salt water (Howes et al., 1984; Capone and Kiene, 1988). Sulfate reduction could lead to the loss of peat which had been deposited while freshwater conditions prevailed. This would in turn lead to an apparent increase in water level (sea level rise + compaction of peat due to further decomposition). The magnitude of this process and rate at which it would occur are unknown and should be the subject of research in the near future.

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SECT. 4. TIDAL RANGE EFFECTS

The Effect of Sea Level Rise on Tides in Estuaries and Bays

The tidal motion in estuaries and bays is forced by the ocean tide at the open boundary. Therefore, in addition to the increase in depth, the effect of sea level rise on the tide in the estuary enters through the ocean boundary tide. The report suggests that sea level rise will be accompanied by an increase in tidal range in the ocean. This is not necessarily true, as the response of the water masses in the ocean and coastal seas to the various frequencies contained in the tidal forcing strongly depends on the resonance frequency. Without carrying out more detailed calculations, it will be difficult to make a definite statement whether the offshore tidal range will increase or decrease with a rise in sea level.

Preliminary calculations pertaining to co-oscillating undamped tides and presented in the report suggest that sea level rise can considerably alter the response of the estuary to those ocean tide components having frequencies close to the natural frequency of the estuary. These calculations assume zero friction, an assumption which for most estuaries is not justified. When including friction, it is expected that the effect of sea level rise on the tide will be less dramatic than the effects predicted by the zero friction models.

In evaluating the effects of sea level rise on tides and tidal circulation, more attention should be given to the combined effect of a rise in still water in the estuary and a change in ocean tidal amplitude. In particular, the change in the ratio tidal amplitude/depth is important. It is the magnitude of this ratio that determines the importance of the "nonlinear effects," including superelevation, residual circulation, and deformation of the tide.

Research Needs

To evaluate the effect of sea level rise on tides and tide-induced circulation, I suggest that, for a number of sea level rise scenarios, including changes in the offshore tide,

- o calculations be carried out for a few schematized estuaries, using analytical techniques and field equations that include friction;
- o calculations be carried out for a number of real estuaries using calibrated and verified hydrodynamic models (San Francisco Bay, Tampa Bay, Biscayne Bay, etc.).

SECT. 6. INTERACTION WITH NATURAL FEATURES AND CONSTRUCTED WORKS

Natural Features

Natural Inlets: An increase in sea level increases the cross-sectional area and the bay tidal amplitude. Assuming a uniformly fluctuating bay level the effect of these changes on the inlet velocity follows from

$$\hat{u} = \frac{2\pi A_b a_b (1 + \frac{\alpha}{ab})}{T A_c (1 + \frac{\epsilon}{A_c})} \qquad (A.5)$$

where $\hat{u}$ is the amplitude of inlet current velocity, A_b is the surface area, T is tidal period, A_c is inlet cross-sectional area, a_b is bay tidal amplitude, and α and ϵ are the increase in respectively the bay tidal amplitude and the cross-sectional area. If $\alpha/a_b < \epsilon/A_c$ the inlet velocity decreases, if $\alpha/a_b > \epsilon/A_c$ the inlet velocity increases.

Stability of inlets can be based on

$$A_c = aP^m \qquad (A.6)$$

In evaluating the effect of sea level rise on inlet stability it should be realized that not only P but also a and m are affected by sea level.

the tide in the bay is little affected by a rise in sea level, it is expected that the inlet cross-sectional areas will adjust to the rising sea levels and retain the original cross-sectional area.

Increased sea level will lead to more frequent breaching of barrier islands. The newly created inlets will affect the existing ones.

Constructed Works

Incorporate expected increase in sea level in the design of structures. Many coastal structures are designed for storm conditions. When restricting attention to the lifespan of a structure, isn't sea level rise negligible with regard to the uncertainties in the level of storm surges? Carry out sensitivity analysis.

Research Needs

Using the present techniques to analyze the stability of inlets, evaluate the effect of sea level rise for a few typical inlets. Include breaching of barrier islands.

SECT. 9. UPRIVER SALTWATER PENETRATION

Estuarine Transport Processes/Salt Intrusion

The upestuary transport of dissolved constituents, including salt, is the result of complex interactions of residual currents and mixing. The complexity of the processes in many cases defies our intuition. For example, increasing the tidal range does not necessarily imply an increase in salt penetration. Abraham et al. (1986) showed that in one of the two branches of the Nieuwe Waterway (Rotterdam) maximum salt intrusion occurred during neap tides and in the other during average tide conditions.

Residual currents and mixing processes in an estuary are determined by the external forcings

- wind,
- ocean tide off the estuary mouth, and
- density gradients in the estuary

and by the

- bathymetry of the basin.

A rise in sea level results in an increase in depth and an increase/decrease in the ocean tidal range and indirectly through the mixing processes affects the density gradients. We have a reasonable understanding as to how external forcings and bathymetry affect residual currents and mixing processes in a qualitative sense. For quantitative realistic estimates, our present knowledge of especially turbulent exchange processes of mass and momentum in the presence of stratification is too limited. Therefore, any estimate of the effect of sea level rise on transport processes/salt intrusion in an estuary has to be of a qualitative nature!

Following are a few examples as to how residual currents and mixing are affected.

Classical estuarine circulation (= gravitational circulation).

$$u = \frac{g}{\rho} \frac{\partial \rho}{\partial x} \frac{h^3}{N_z} \tag{A.7}$$

where u = velocity, g = gravity acceleration, ρ = density of water, x = longitudinal coordinate, h = depth, and N_z = coefficient for turbulent vertical exchange of momentum. Note the dependence of u on h^3. The longitudinal dispersion coefficient, D, associated with gravitational circulation is

$$D :: \left(\frac{g}{\rho} \frac{\partial \rho}{\partial x} \right)^2 \frac{1}{N_z^2} \frac{1}{K_z} h^5 \tag{A.8}$$

where K_z = coefficient for turbulent vertical exchange of mass. Note the dependence of D on h^5.

Horizontal residual currents driven by longitudinal density gradients:

$$u :: \frac{\bar{h}^2}{U_T} \tag{A.9}$$

where U_T = amplitude of tidal velocity. Overbar denotes averaging on the cross-section.

Residual currents (in a coastal channel) induced by boundary conditions:

$$u :: \left(\frac{a}{h} \right)^2 \qquad \text{(A.10)}$$

where a = amplitude of ocean tide.

The foregoing examples show the strong dependence of residual currents and mixing (dispersion) on the water depth, tidal amplitude, and tidal current. Obviously the effect of sea level rise on transport processes cannot be ignored.

Research Needs

I have taken this to imply evaluating the effect of sea level rise on transport processes using the present state of the art of physical processes in estuaries. My suggestion is for a few sea level rise scenarios (i.e., increase in water level and increase/decrease in ocean tidal amplitude) and for estuaries with different dominant transport processes to calculate the change in salt intrusion. Calculations to be carried out for a few schematized estuaries using analytical techniques. This will elucidate the role of changes in depth and ocean tidal amplitude. Using existing calibrated and verified numerical models, carry out calculations using the actual bathymetry of the estuary.

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PAST AND FUTURE SEA LEVEL RISE IN THE NETHERLANDS

Introduction

In the first part of this paper a view will be given of the changes along the Dutch coast in mean sea level and coastal erosion during the last century (centuries). The second part will deal with the expected impacts for the Netherlands of an accelerating sea level rise.

Past Changes in Mean Sea Level and Coastal Erosion/Accretion

The longest and oldest set of mean sea level in the world is recorded in the Netherlands in Amsterdam (Van Veen, 1954). Fig. A.2 shows the filtered data sets of Amsterdam, Brest, and Den Helder together with the global average data set made by Gornitz et al. (1982). The filter process used cubic spline functions, where these splines were fitted to the data with a least squares method. The curves of Amsterdam and Brest show a remarkable "bend" between 1850 and 1900, which also seems to be present in the curves of Den Helder and the global average. It is likely that sea level rise was not constant during the last 300 years and even negligible before 1850.

The same filter procedure was used for eight stations along the Dutch coast (Fig. A.3). A map showing the positions of these stations is given in Fig. A.4. The curves of the eight stations agree rather well after 1900, but before 1900 only Flushing and IJmuiden show a large fluctuation, which is not or only hardly present in the other curves. An explanation of this behavior has not yet been found. For comparison, the trends have been calculated over the period after 1900 to avoid the differences before 1900. Table A.5 shows the mean sea level rise together with the mean rise of

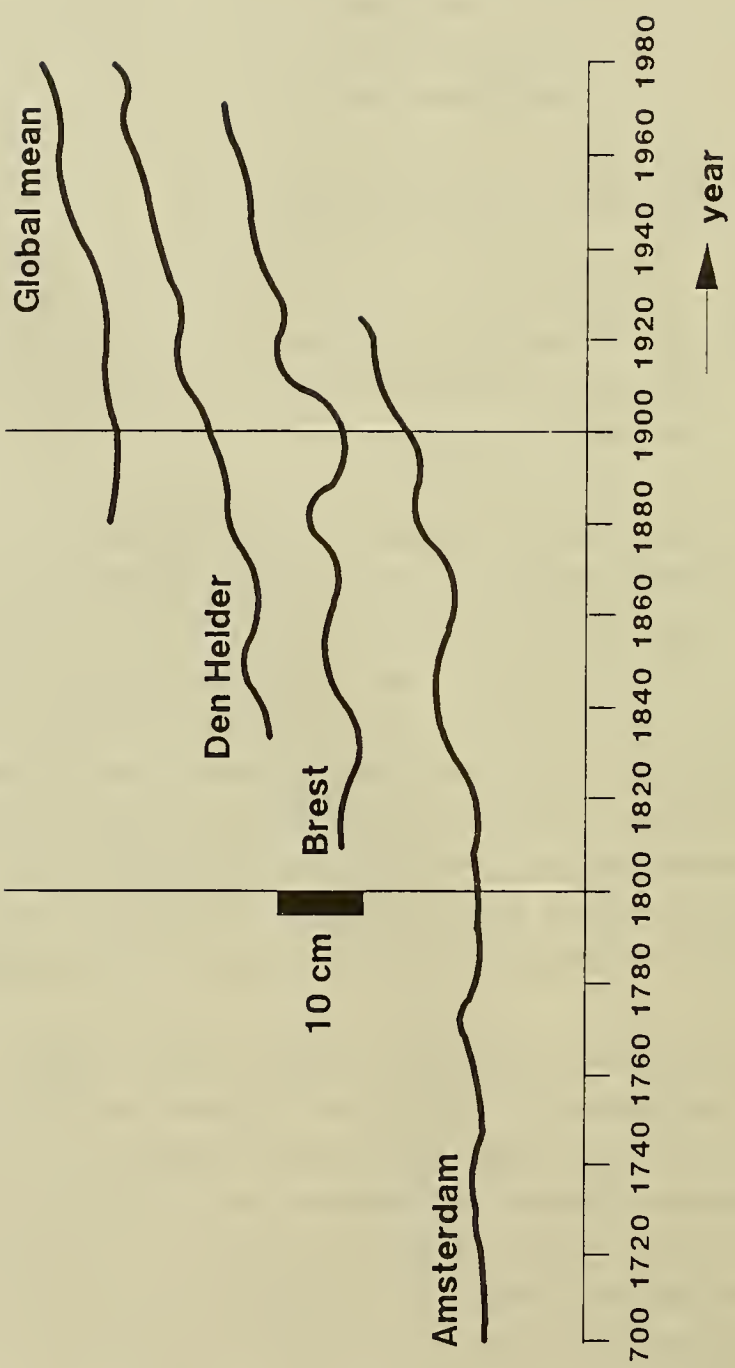


Fig. A.2. Filtered Mean Sea Levels at Amsterdam, Brest, and Den Helder, and Global Mean.

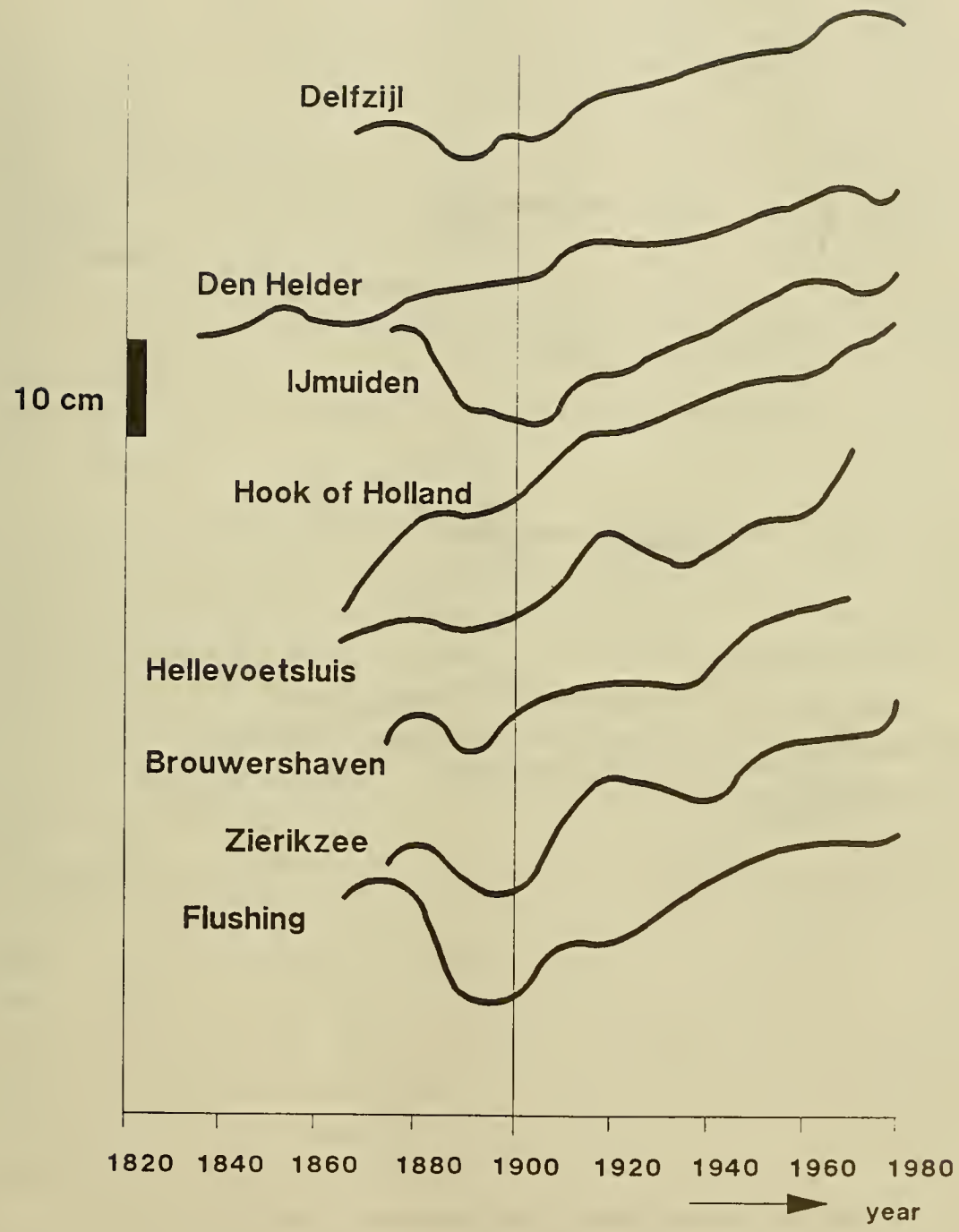


Fig. A.3. Filtered Mean Sea Levels of the Main Dutch Gages.

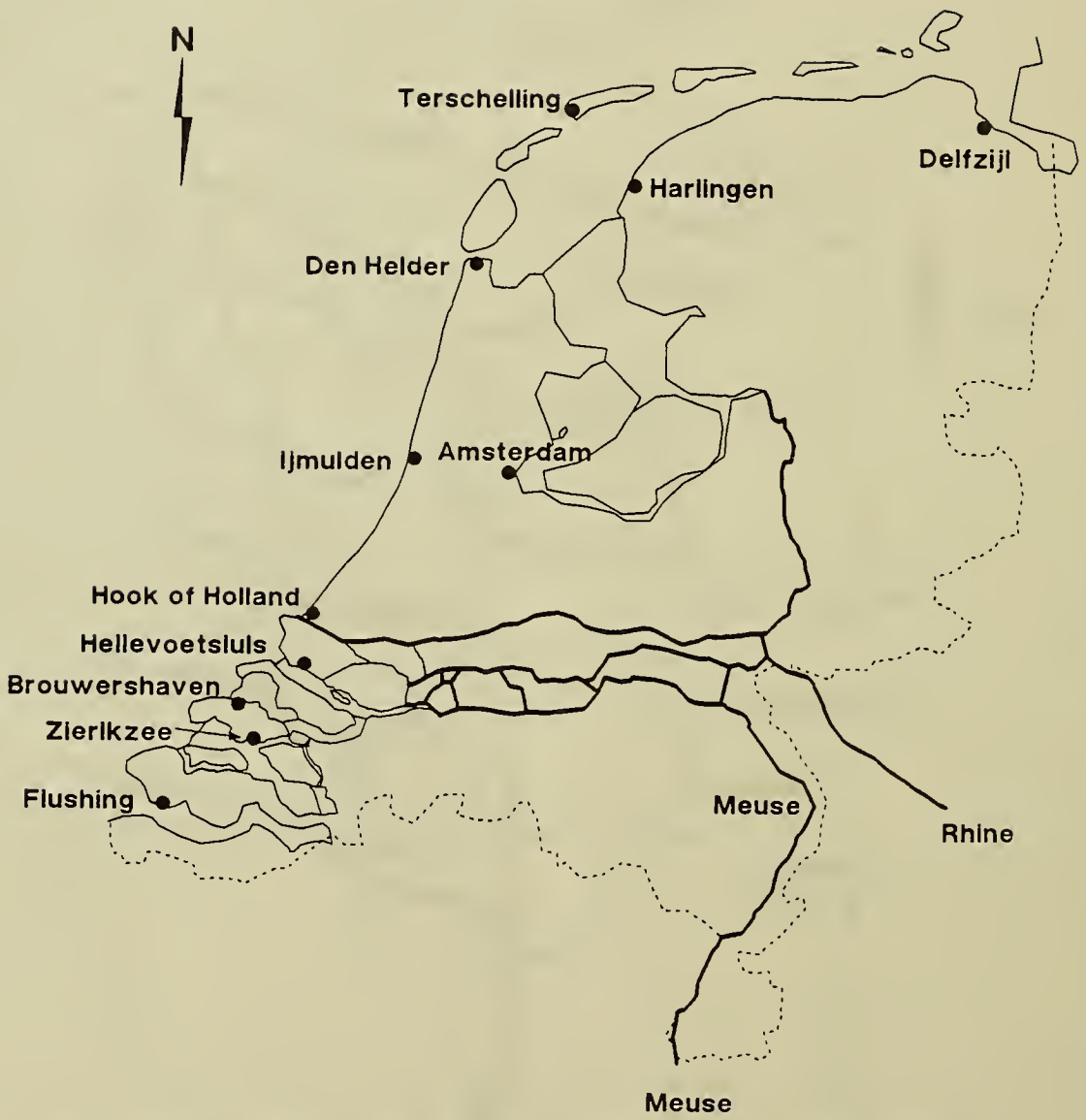


Fig. A.4. Map Showing the Main Dutch Gages.

Table A.5. Mean Sea Level Rise and Mean Increase of High Water, Low Water, and Tidal Range in cm per Century Over the Period 1901-1986

	Sea Level	High Water	Low Water	Tidal Range
Flushing	21	33	18	15
Hook of Holland	20	-*	-*	-*
IJmuiden	22	28	18	9
Den Helder	14	19**	8**	11**
Harlingen	11	30**	8**	21**
Terschelling	10	20**	-14**	34**
Delfzijl	17	-*	-*	-*

*No regular trend.
**Over the period 1933-1986, because of the closure of the Zuider Sea in 1932.

the high water level, the low water level, and the tidal range. The relative mean sea level rise along the Dutch coast fluctuates between 10 and 22 cm per century. In the north of Holland the mean sea level rise is less than in the south, which agrees well with the known tilting of this region, due to the glacial rebound of Scandinavia.

It can be seen from Table A.5 that not only the mean sea level but also mean high and low water and mean tidal range are changing. So for our dikes we have to deal with a rise of 19 to 33 cm per century (rise of high water) instead of the 10 to 22 cm rise of mean sea level per century. The different rise of mean high and low water is due to an increasing tidal range. The increase of tidal range can partly be explained by dredging and harbour works. Due to sea level rise the tidal range changes, but only in a minor way (less than 3 cm per century, assuming 20 cm sea level rise per century); see also the next section. This means that a part of the increase of the tidal range cannot be accounted for with our present knowledge.

It is not surprising that the Dutch coast as a whole has been eroded due to the sea level rise and the increase of tidal range, shown in the previous section. This does not mean that the coast is eroding everywhere; at some locations the coast is even moving seaward. Places along the coast

protected by dikes or sea walls do not retreat, due to severe maintenance. The erosion and accretion rates are the smallest (less than 1 m per year) along the central part of the Dutch coast and the largest (more than 10 m per year) in the north along the Wadden Islands (Figs. A.5, A.6, and A.7) (Kohsiek, 1988). In the figures three bar-graphs are given; the first one gives the trend of erosion/accretion of the low-water line over the last 100 years, the second one shows the erosion/accretion also of the low-water line over the last 20 years, and the third one the trend of erosion/accretion of the -4 m line over the last 20 years.

The northern part (the Wadden Islands) shows great differences both in space and in time of the erosion/accretion rates. The island of Texel is eroding along its whole coast from 2 to 11 m per year as well during the last 100 as during the last 20 years and both for the low-water line and the -4 m line. The other Wadden islands show variable results (Fig. A.5).

The central part has an eroding coast in the south and the north and an accreting coast in the middle. The two low-water line graphs representing the last 100 and 20 years show only small differences. This is not the case for the two graphs of the low-water line and the -4 m line, where the differences are remarkable (Fig. A.6). Partly these differences are due to the groins, causing the erosion of the low-water line to be small or even zero while the shore face (the -4 m line) retreats much faster. It is probable that the retreat of the shore face is already showing the future retreat of the low-water line, which is delayed by the groins.

In the southern part (the Delta Islands) coastal development is very variable; most parts show erosion, some are stable, and some accrete. The bar graphs of the low-water line for the last 100 and 20 years show great differences due to the Delta works, which influenced the changes of the last 20 years significantly (Fig. A.7). The two graphs with the trends of the low-water line and the -4 m line over the last 20 years show good agreement.

During the past 20 years the Dutch coast has lost 5.3 million cubic meters of sand per year along the eroding parts and gained 3.4 million cubic meters per year along the accreting parts. As a whole, the Dutch coast is losing 1.9 million cubic meters per year.

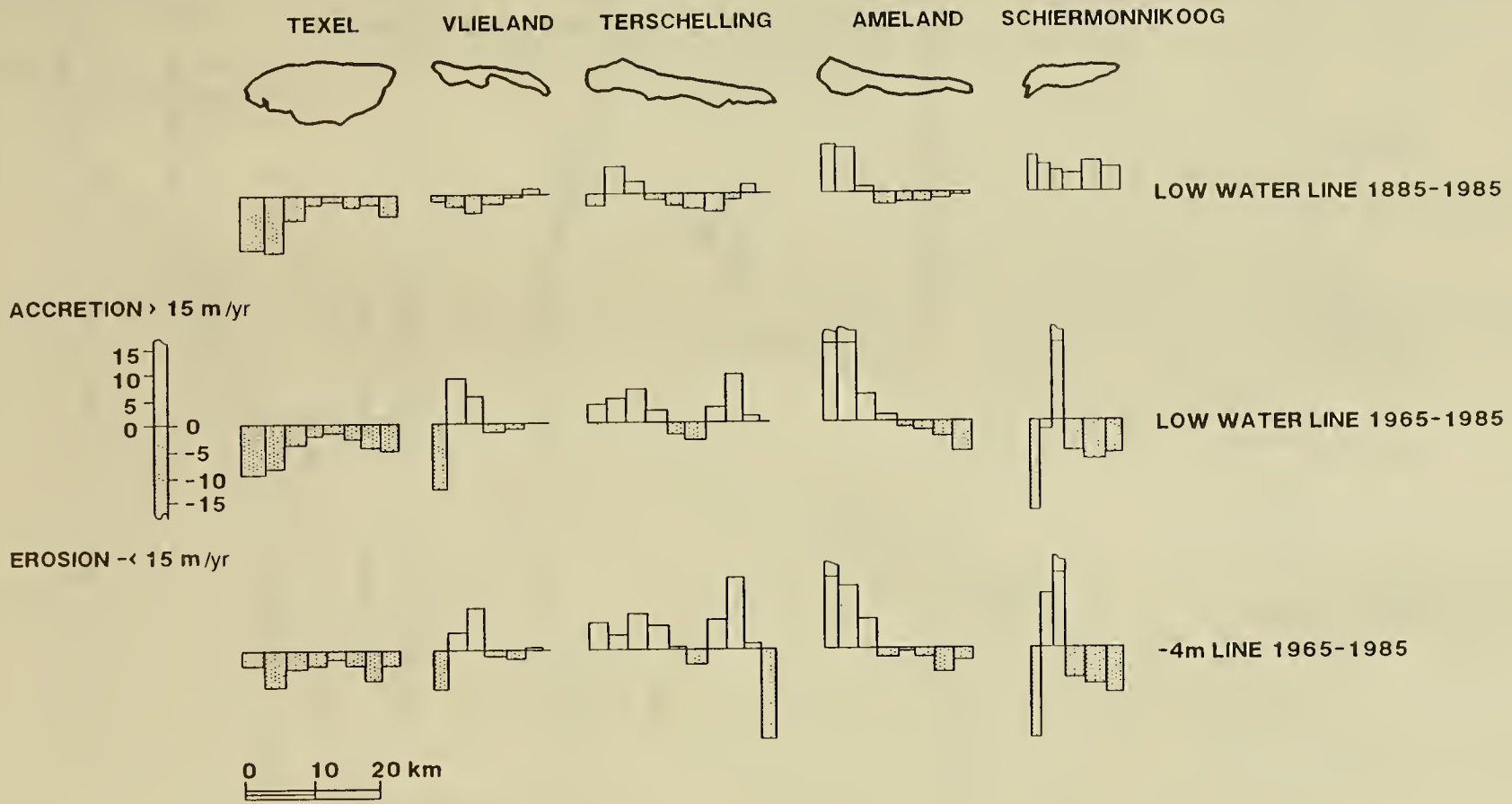


Fig. A.5. Erosion and accretion of the Wadden Islands.

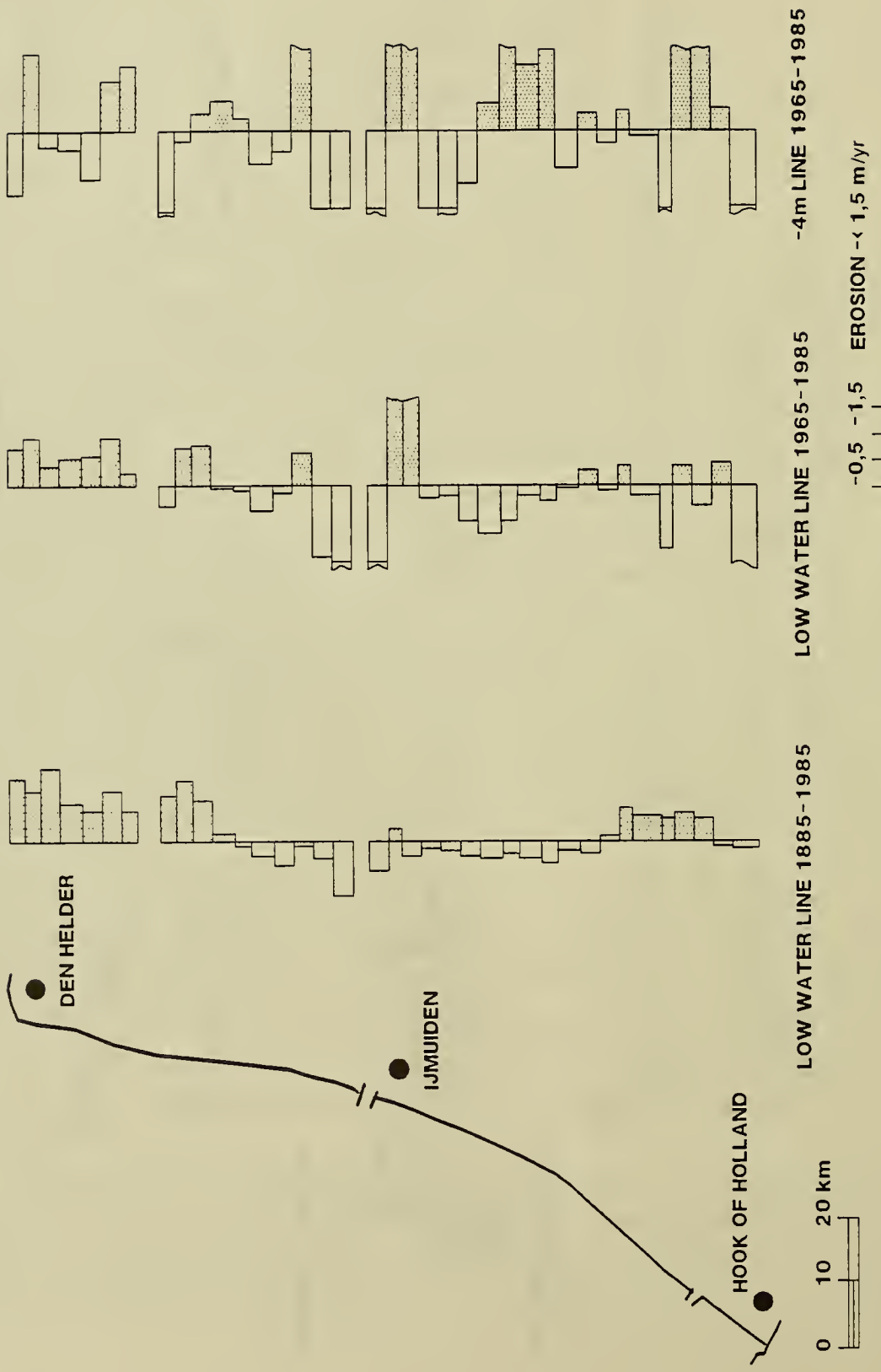


Fig. A.6. Erosion and Accretion of the Central Coast.

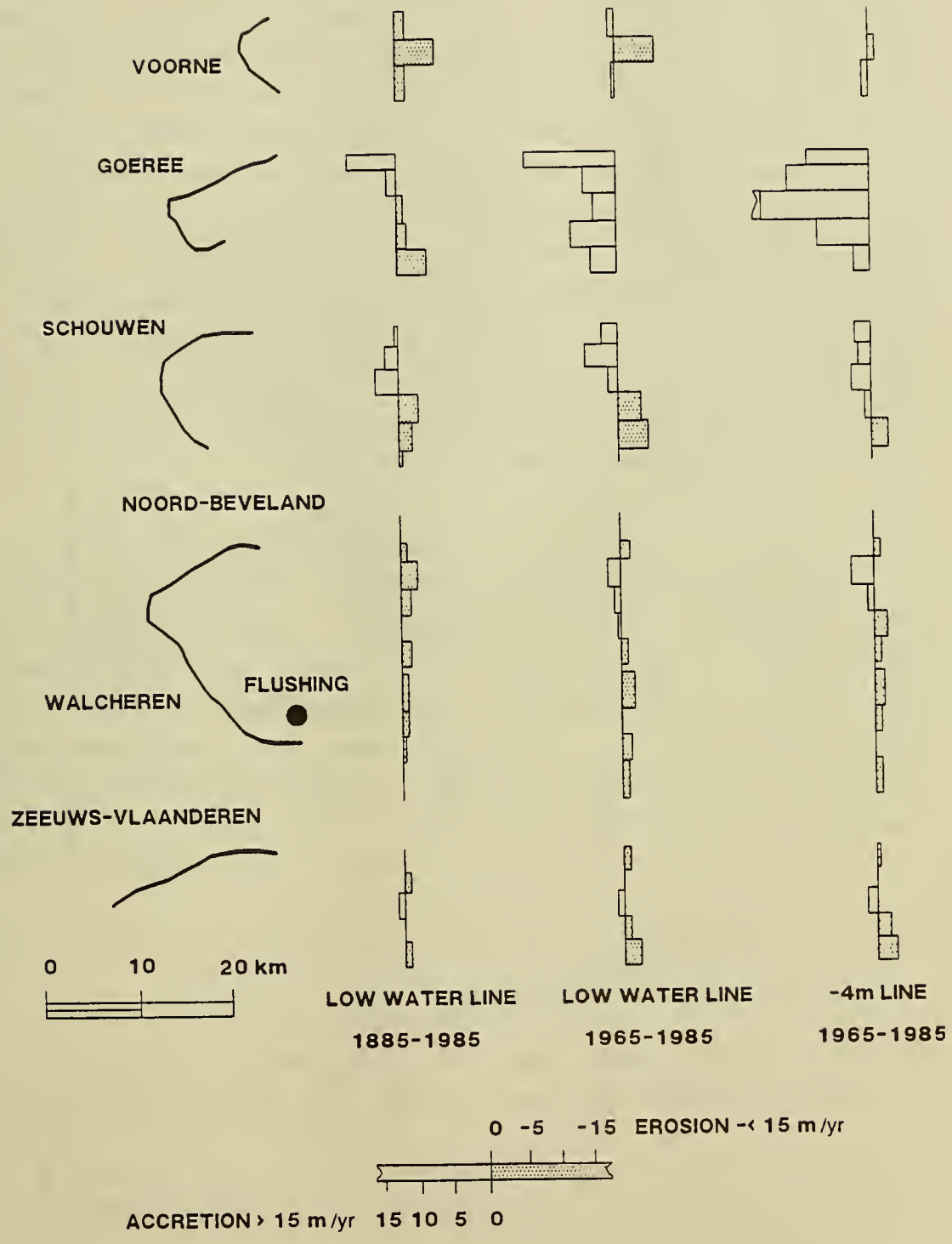


Fig. A.7. Erosion and Accretion of the Delta Islands.

It is clear that one must be careful with the interpretation of short time series of coastal erosion and accretion. This is especially the case when the phenomenon of "Sand Waves" is important along a certain part of the coast. Along the Dutch coast there are major sand waves moving along the southern and the northern part; along the central part the sand waves are rather small (Kohsiek, 1988). The sand waves migrate to the north in the southern and central part and to the east in the northern part. The amplitudes on the Wadden Islands vary between 10 and 1200 m, while the migration speed varies between 220 and 450 m per year, e.g., the sand waves moving along the northwestern part of the island Schiermonnikoog have a period of about 30 years and an amplitude of about 300 m (Fig. A.8). Along the central part, amplitudes are 25 to 50 m and migration speed is 30 to 100 m per year. In the southern part of the Dutch coast, the sand waves have amplitudes between 25 and 200 m and migrate 30 to 300 m per year.

Impacts of Sea Level Rise on the Netherlands

Due to the greenhouse effect, mean sea level along the Dutch coast will rise more rapidly in the coming century than in the recent past. In this presentation a certain rise (in most case 1 meter) in sea level over the next 100 years is chosen in order to investigate the possible impacts of future sea level rise in the Netherlands. The results presented herein are a summary of a Dutch report entitled "Zeespiegelrijzing - Worstelen met wassend water" (de Ronde et al., 1986).

A sea level rise of 1 meter or more has serious implications for:

* Safety of dikes and other defense structures along the coast and the lower parts of the main rivers.

* Morphology of dunes, shore face, and estuarine systems. A critical issue is whether the bed can rise as fast as the sea level rises.

* Stability of ecosystems. How will vegetation and animal life change with sea level rise? The Wadden Sea is of great importance for North Sea fish and migratory birds.

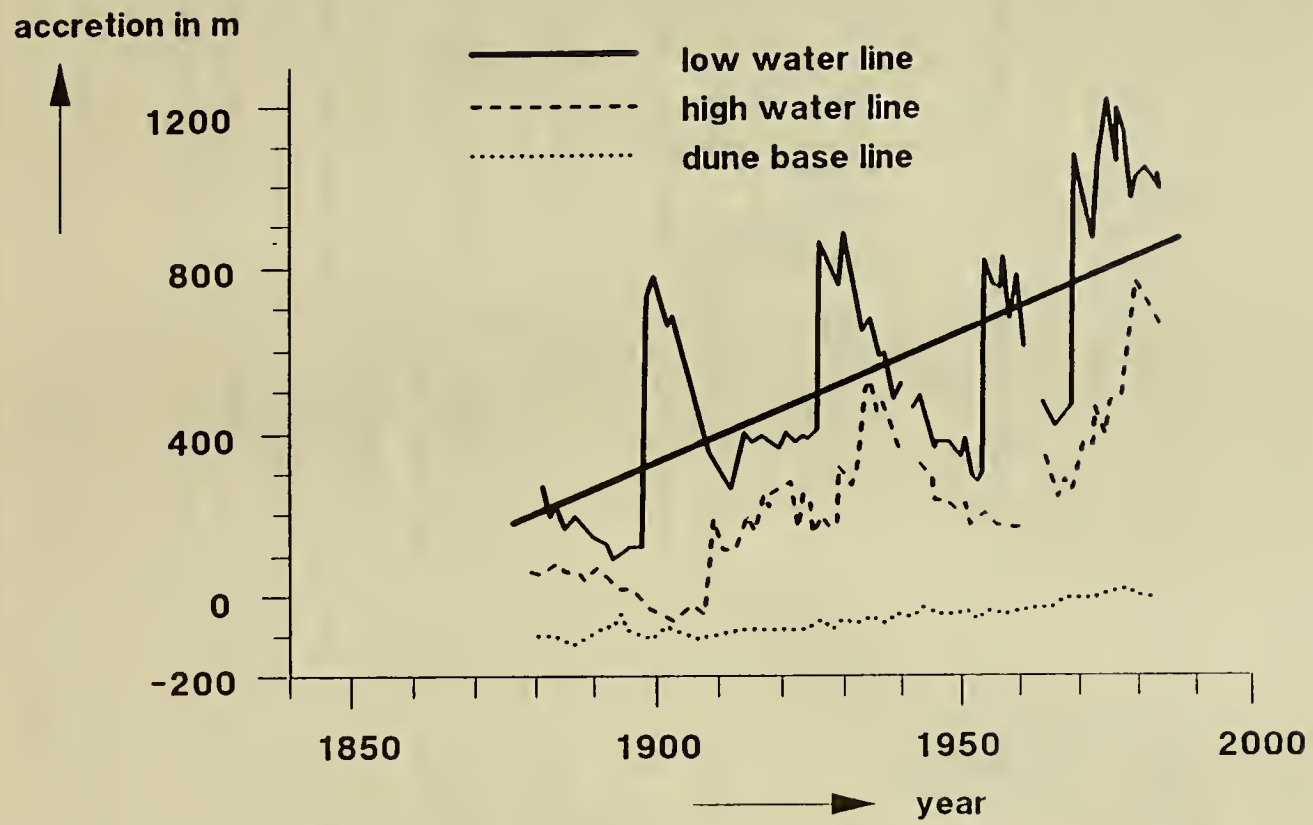


Fig. A.8. Sand Waves on the Wadden Island Schiermonnikoog.

* Water resources. Increasing seepage of salt water into inland polders and increasing salt intrusion via the rivers are of great concern. The present natural drainage of the IJssel Lake during low tide may have to be replaced by a pumping system. Storage of drinking water below the dunes may be diminished.

* Other structures. Harbours, bridges, locks, etc. may fact alteration.

Preliminary impact studies by Rijkswaterstaat were first conducted with an extreme scenario of a 5 m sea level rise in order to gain a better understanding of the major mechanisms involved. The results obtained will be discussed in the following sequence: impact on the tidal system in the North Sea, on wage propagation, on coastal morphology, and on ecology. Finally, the costs involved will be discussed.

Impact on the Tidal System in the North Sea - A two-dimensional hydraulic computer model (WAQUA) of the North Sea and part of the Continental Shelf (Verboom et al., 1989; see Fig. A.9) was used to calculate tidal system changes compared to the present situation for sea level scenarios of 2.5 m and 5 m higher, and 2.5 m lower, than today (de Ronde and de Ruijter, 1986).

The main effect was that at higher sea levels and greater water depths the tidal wave propagated faster and tidal amplitudes (water levels and currents) increased. A second effect was the changing of the positions of the amphidromic points (=nearly zero amplitude) (Fig. A.10). Thus, a location from which the amphidromic point moves away will experience a higher amplitude, and the location to which the amphidromic point moves will experience a lower amplitude. These two effects combined resulted along most North Sea coastlines in an increase of the tidal amplitude (Fig. A.11). Only locations near the line Cromer-Den Helder experienced a decrease as the southern amphidromic point drew nearer. Here the decrease of the tidal amplitude due to the moving away of the amphidromic is stronger than the increase due to the greater water depth.

With regard to storm surges, model results showed (de Ronde and de Ruijter, 1986) that a rise of the sea level of 5 meters, while wind

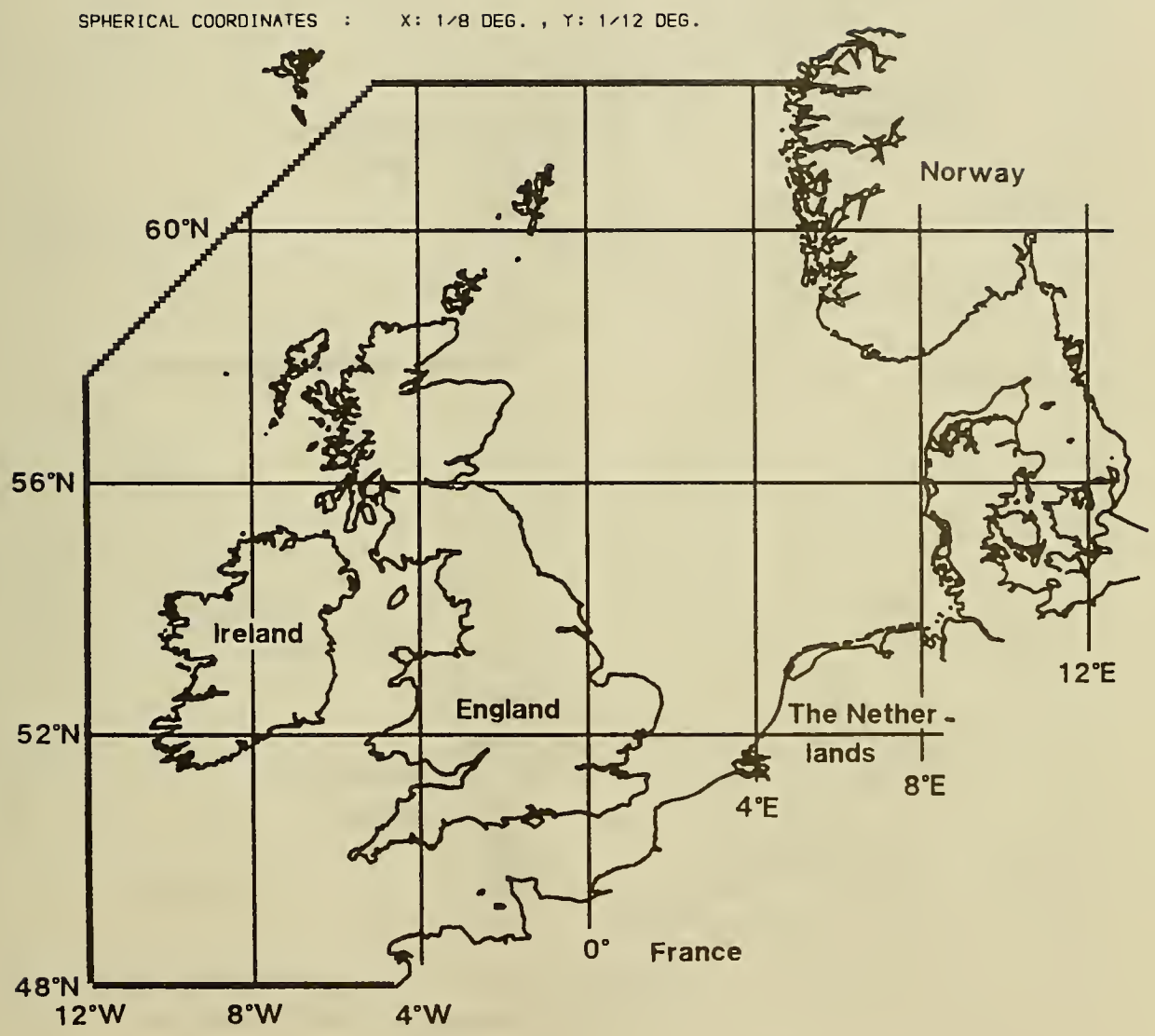


Fig. A.9. The Outlines of the Continental Shelf Model.

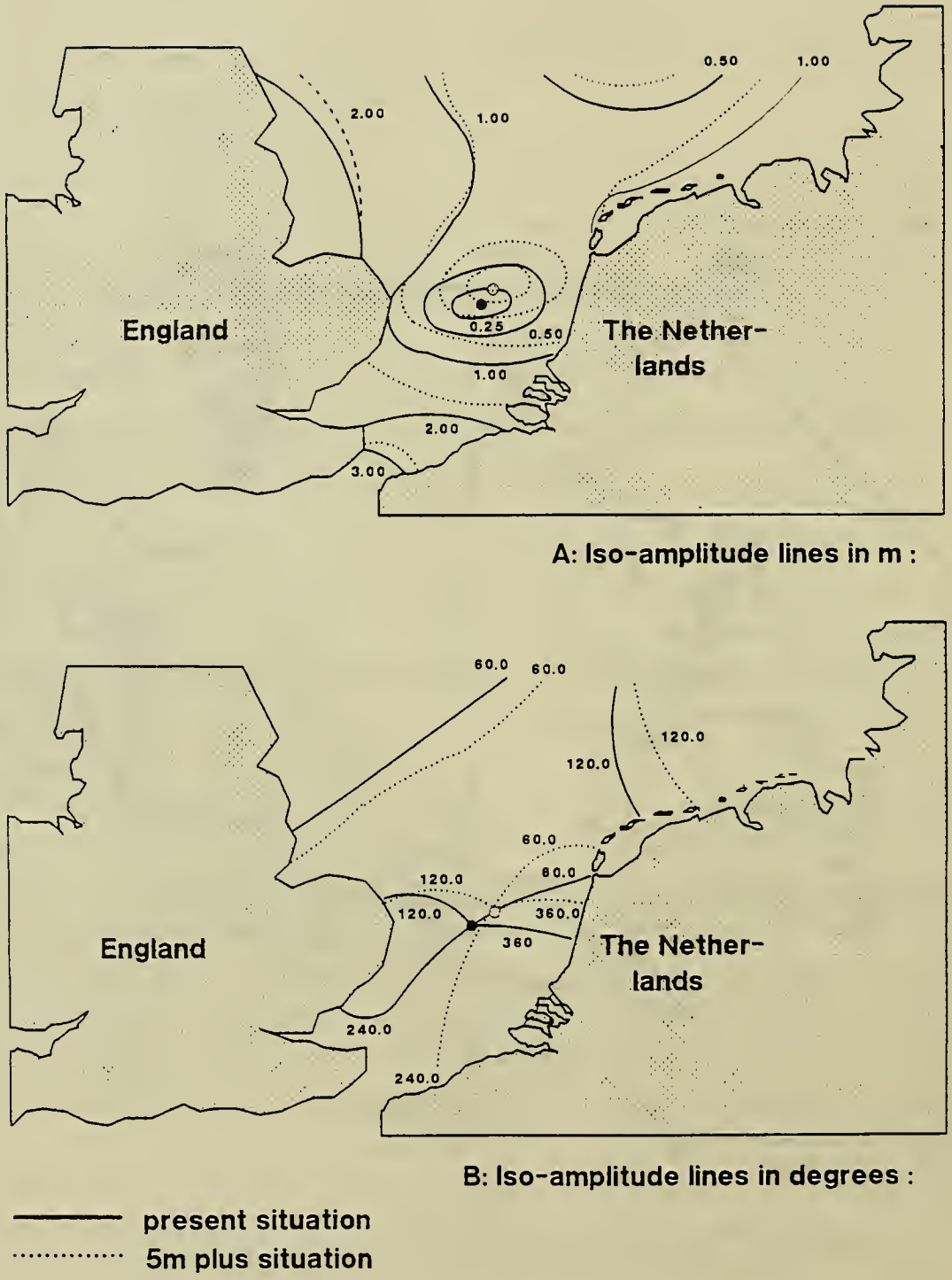


Fig. A.10. The Change in the Semidiurnal Amphidrome in the Southern North Sea, with a Sea Level Rise of 5M: a) Iso-Amplitude Lines in Meters, b) Iso-Amplitude Lines in Degrees.

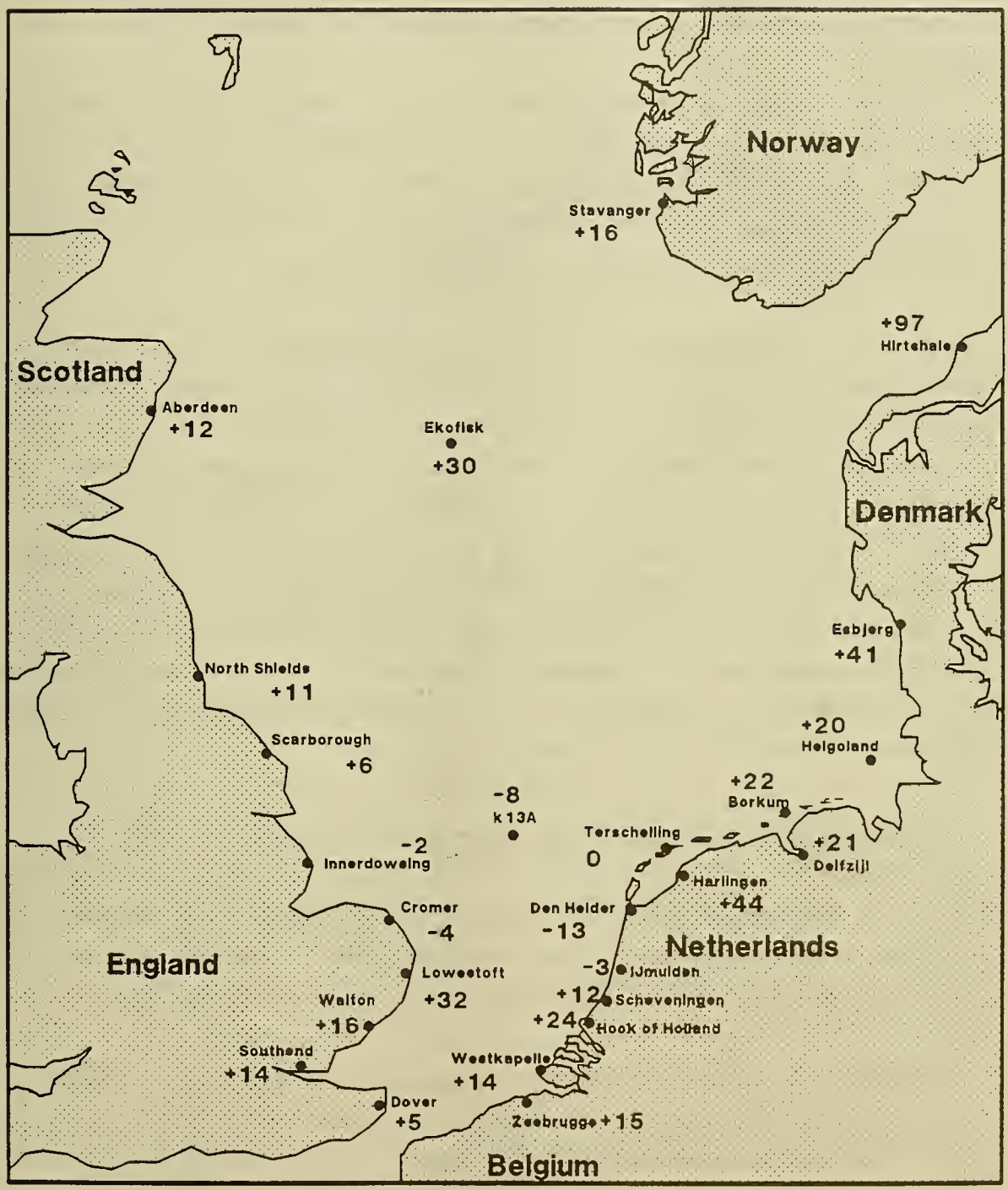


Fig. A.11. Percent Change in Tidal Amplitude, with a Sea Level Rise of 5 m.

velocities remained unchanged, caused a change of the storm surge heights between -20 and +10 cm. The effect of sea level rise on storm surges is thus very small, and the possible increase or decrease in the number of severity of storms will be much more important for changes in storm surge heights.

Another feature which may be important is a change in the residual transports, which will have impacts on the morphology and the marine ecosystem, especially in the Wadden Sea. Model results show a change in the residual transports when sea level rises by 1 meter (Fig. A.12), but results are still too inaccurate for quantitative use.

Impact on Waves and Swell - A higher sea level and a greater water depth caused less dissipation of waves and swell. Wave heights increased, especially in shallow water. Many of the smaller waves, which break and dissipate on shallow sandbanks in the present situation, will continue unbroken in the case of increased water depth. For instance, calculations of the wave height with the HISWA wave model near the Haringvliet sluices, a location behind a shallow sandridge called the Hinderplaat, showed an increase in wave height from 90 cm to 210 cm when the sea level rose 5 m over an unchanged sea bed. However, if the sea bed were to rise as fast as the sea level, the wave height would hardly change. In reality the outcome will depend on the change of the morphology.

Impact on the Morphology - The Netherlands coast can be roughly divided into three different parts: the Delta area in the South with the Eastern and Western Scheldt; the central coast between Hook of Holland and Den Helder without important inlets; and the Wadden Sea area with the inlets between the islands. The easiest part to study is the central coast.

The coastal defense of the central coast consists mainly of dunes. The present situation is close to stable, but a sea level rise threatens to destroy this delicate balance and will cause the coastline to retreat. The present beach slope is about 1:60, so a rise of 1 meter may result in a retreat of 60 m. A complicating factor is that the dunes are composed of finer grain sand than that found on the beach and shore face. The expected

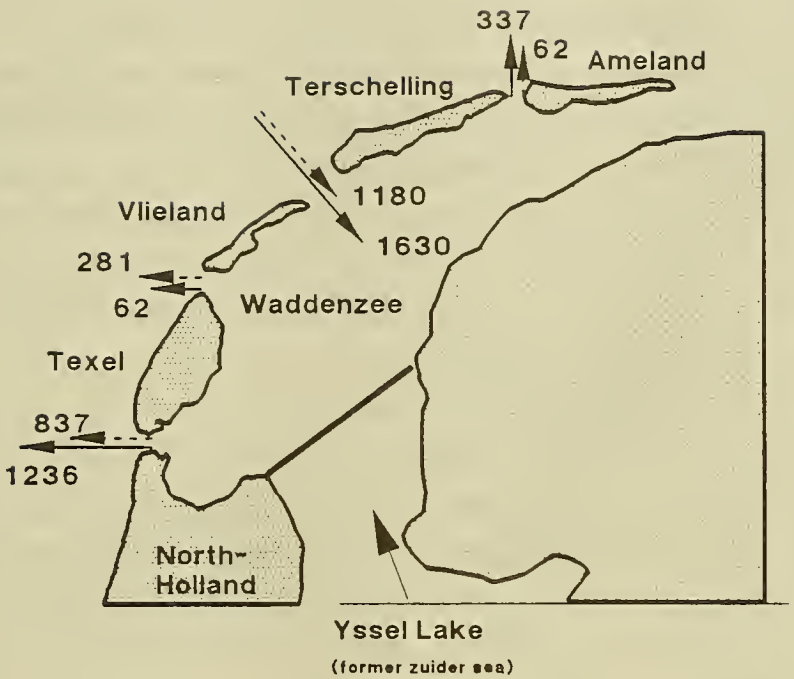


Fig. A.12. The Change in the Residual Transports in the Wadden Sea, with a Sea Level Rise of 5 m.

slope of a beach consisting of this finer sand will be less than 1:60 and consequently the retreat of the coastline may be as great as 80 to 150 m. After such a retreat the dunes may not be strong enough to withstand a design storm surge (design frequency 10^{-4}), resulting in the need to strengthen parts of the dunes. A retreat of dunes and coastline would be undesirable in many cases, due to the presence of harbours, other constructions, and valuable dune areas, and so for large parts of the Netherlands coast measures such as sand replenishment may be necessary. For the coastal stretch protected by dikes (such as the Hondsbossche Zeewering) strengthening is the only solution.

The morphology of the Wadden Sea is very complex. Even the present situation is difficult to model. Added to this uncertainty is the fact that the Wadden Sea has still not attained equilibrium since the closure of the Zuider Sea in 1932 (now IJssel Lake).

Presently there is a transport of sand from the North Sea into the Wadden Sea of 30 x 10^6 m^3/year (standard deviation 15 x 10^6). If sea level rises 1 meter in 100 years, an addition of 28 x 10^6 m^3 sand per year is needed for the sea bed to keep pace. So for the area at large, bottom rise can probably keep up with this rate of sea level rise. So the intertidal area will stay about the same size. This is very important for the ecology of the area (see below).

It should be stressed once more that the morphology of the Wadden Sea is very complex and that the sedimentation theory used here is very simple. For example, of the origin of the present sedimentation (30 x 10^6 m^3/year) only 25% can be traced to the coast of north Holland, the coasts of Texel and Vlieland, and the adjacent North Sea, whereas the origin of the remainder is more diffuse. Even at the present rate of sea level rise it is not known in detail how sedimentation changes the morphology of the Wadden Sea.

In the Eastern and Western Scheldt of the Delta area, erosion will probably occur and the intertidal area will probably decrease. But here, too, it has to be stressed that the processes involved are very complex and predictions difficult to make.

Salt Intrusion - There are two major ways in which salt water can intrude upon the Netherlands (the salt intruding the Netherlands via the pollution of the river Rhine we shall leave aside). The first way is directly from the sea via the river branches coming out into the sea. The Rotterdam Waterway is in this respect the one that causes the main troubles. With the coming of larger and larger ships the Rotterdam Waterway had to be made deeper and deeper. This caused the salt intrusion not only to become more intense, but the salt also intruded further, causing problems at water inlets for agricultural and drinking water purposes. When Europoort, the new harbour area, was finished, further deepening of the Rotterdam Waterway was no longer necessary. The depth could even be decreased.

A rising sea level will again cause an increase of the salt intrusion. Figures A.13 and A.14 show the results of model calculations where sea level was raised over 5 m and where bottom topography remained the same. Big areas such as the Haringvliet, important for fresh water supply, will under the circumstances be intruded with salt water. This salt water will remain in the deeper parts of the Haringvliet and it will be extremely difficult to remove it. If the bottom of the river system rises as fast as the sea level the salt intrusion remains, of course, the same.

The second major way in which salt water intrudes is via groundwater flow through the subsoil (seepage). Large areas of the Netherlands are below sea level, which causes a groundwater flow to these lower parts. Because the deep groundwater layers in the western part of the Netherlands are brackish, this upward seepage is brackish, too. In many polders the seepage rate is more than 0.25 mm per day. If nothing is done grassland and crops would be damaged.

The solution is flushing and rinsing of polders and canals with fresh water. In very dry summers not enough fresh water is available for this purpose and crop damage is unavoidable. The amount of seepage is linearly related to the difference in water level between the sea and the groundwater in the polders. It is also linearly related to the distance between polders and sea and the permeability of the subsoil.

A higher sea level causes a greater gradient and an increase of the seepage. Moreover in certain areas where today there is no seepage, a sea

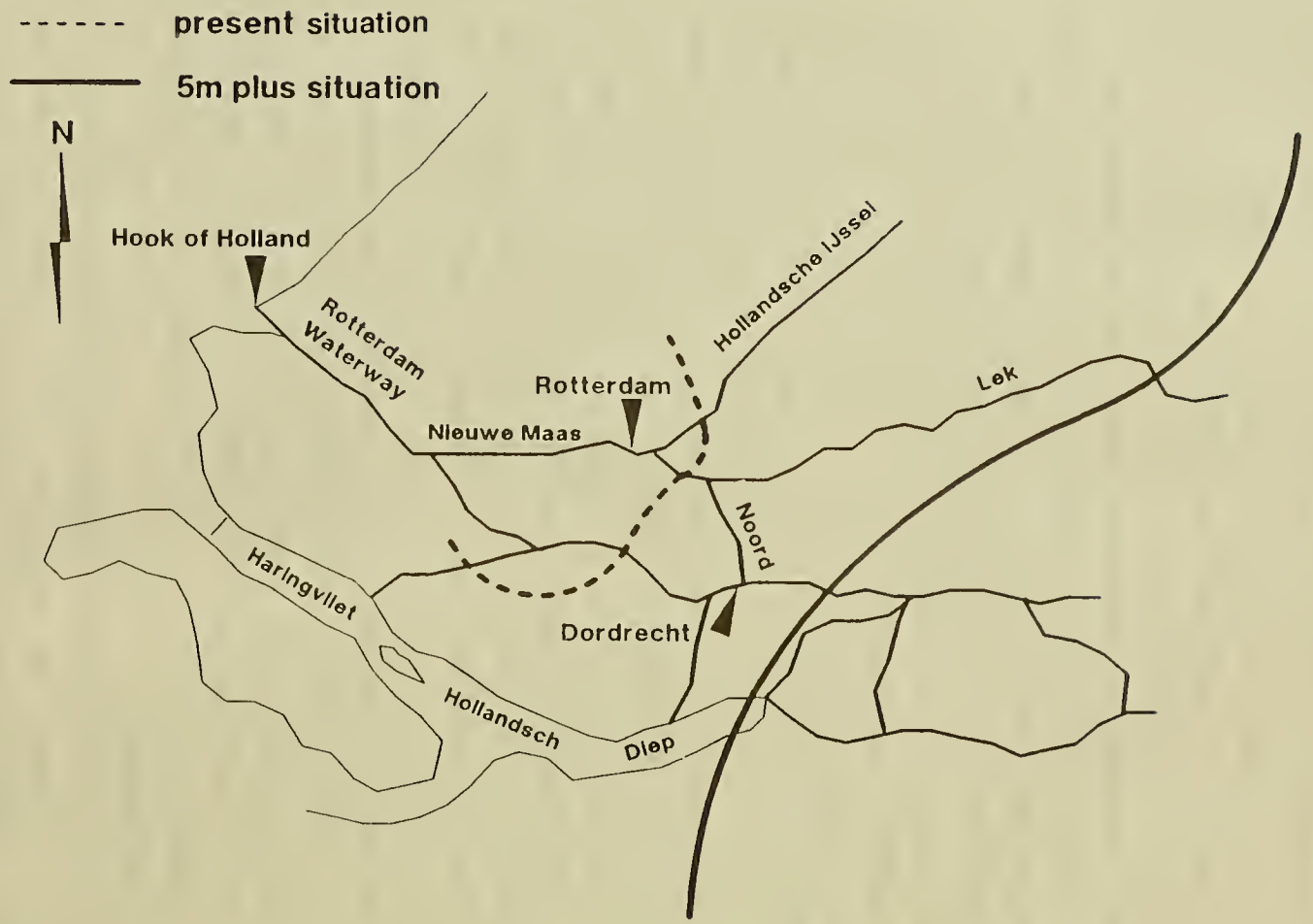


Fig. A.13. The Change in the Salt Intrusion in the Rhine Estuary, with a Sea Level Rise of 5 m.

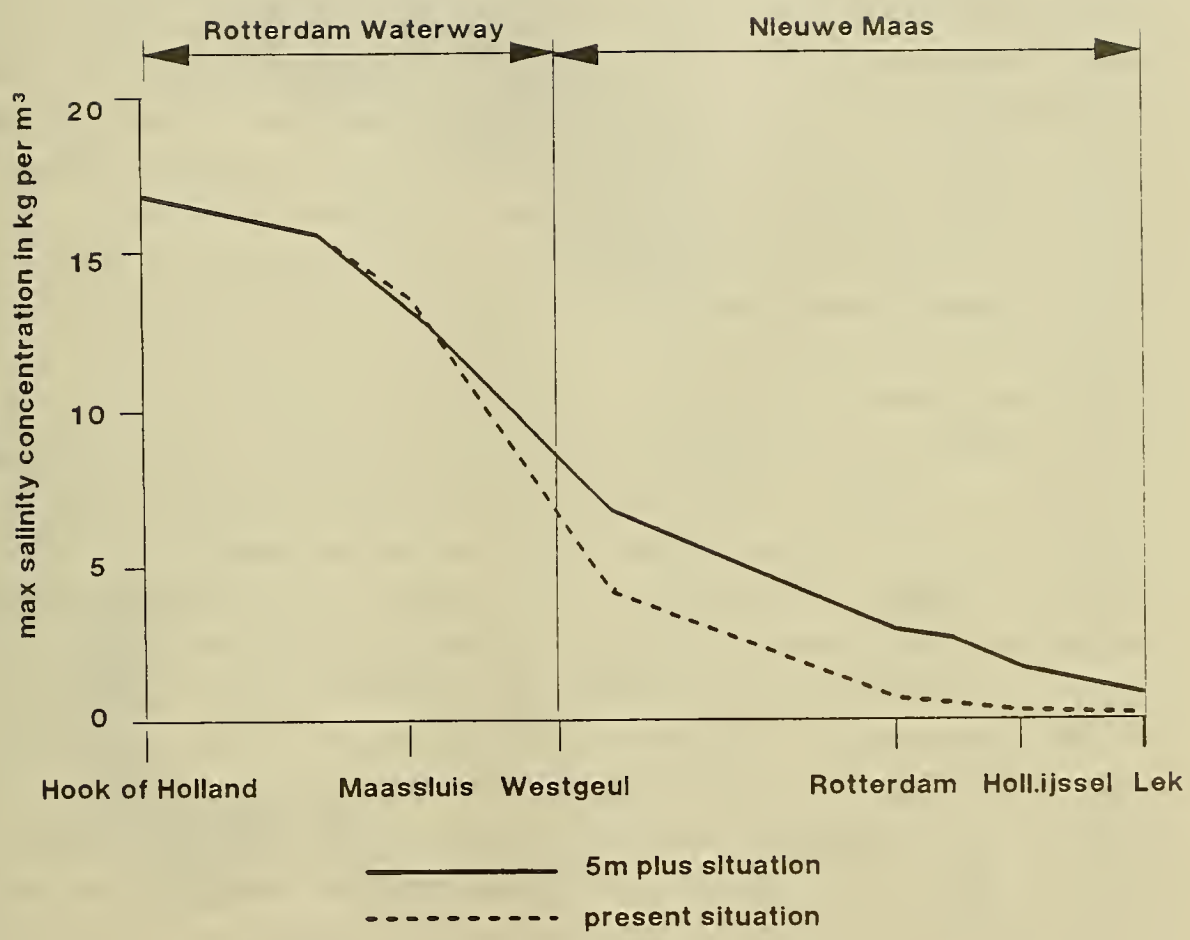
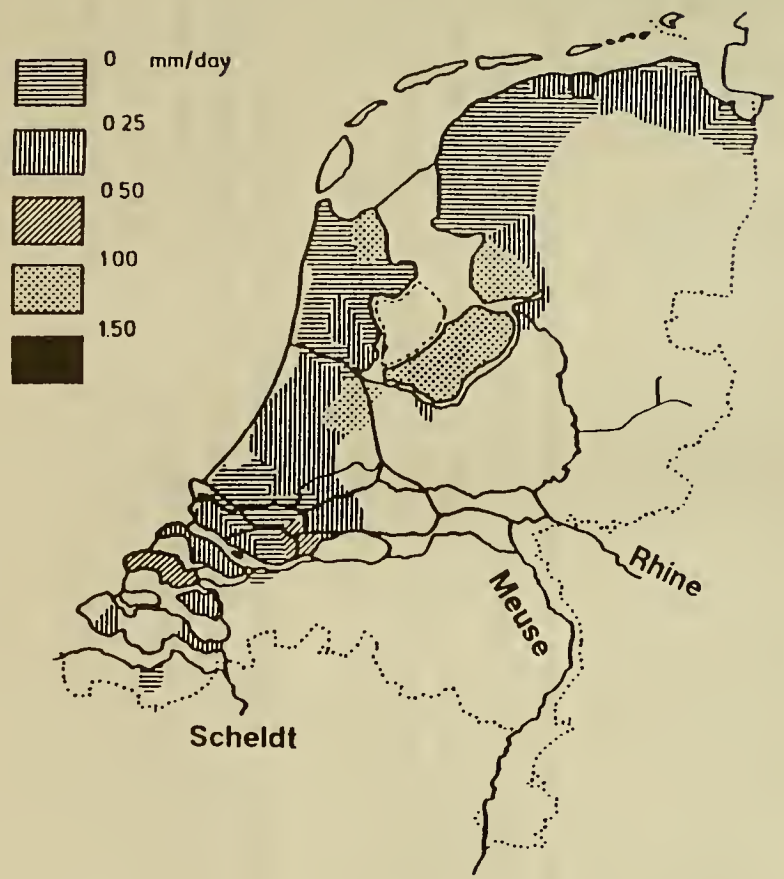


Fig. A.14. The Change in the Maximum Salinity Concentration during a Tidal Cycle in the Rhine Estuary, with a Sea Level Rise of 5 m.

level rise will cause seepage to start. See Fig. A.15 where the present seepage rate is given together with the rate when sea level would be 5 m higher. Simple calculations show an increase of the total seepage of 200 to 300% when sea level rises 5 m. The amount of fresh water needed for flushing of the polders will then increase by about 50%. In dry summers this extra amount is not available.

Impact on Water Management - A main problem for water management has been discussed above, namely the saltwater intrusion. Another important item is the drainage of the surplus of water. This is especially the case during winter when precipitation is large and evaporation is small. Where we are dealing with areas well below sea level there is only one solution for the drainage, namely pumping. In case of a higher sea level all pumping stations have to be rebuilt because the surplus water of polders has to be pumped upwards over a greater height and stronger pumps will be needed. There are also areas that drain naturally by gravity. The solution here seems to be building of new pumping stations when sea level rises. However, in the Netherlands a very extensive area is still drained by gravity, with a very large discharge, where this solution might not be the best one. I am talking of the IJssel Lake and the area (15,000 km^2) that in turn drains into the IJssel Lake. Moreover, the river IJssel comes into the lake and brings about 10% of the discharge of the river Rhine to the lake. The IJssel Lake with an area of 1200 km^2 is drained naturally into the Wadden Sea by two sluices in the enclosing dam. The water level of the lake is kept 50 cm below mean sea level during winter and 20 cm below mean sea level during summer. This makes natural drainage possible during low tide. In a recently performed study it has been found that up to a sea level rise of about 50 cm it is possible to keep on draining the IJssel Lake naturally without raising the water level of the lake. A substantial further increase of the sea level would, for example, make it necessary to build an enormous pumping station with a capacity of 300 x 10^6 m^3/day. This great capacity is needed to keep the water level of the IJssel Lake below 1.30 m (above mean sea level) during extreme discharges of the river Rhine (the design discharge of the river Rhine is 16,500



A: present situation

B: 5m plus situation

Fig. A.15. The Change in the Seepage of Salt Water through the Subsoil in the Netherlands, with a Sea Level Rise of 5 m.

m^3/sec, of which about 1650 m^3/sec goes through the river IJssel into the IJssel Lake).

A possible better solution would be to raise the water level of the IJssel Lake as much as the sea level rises. The consequence is, of course, that all the dikes around the IJssel Lake and partly along the river IJssel must be made higher as well. Moreover, a great number of pumping stations would have to be built along the lake for those areas that drain naturally into the IJssel Lake. Until now only the direct consequences of sea level rise on the water management of the Netherlands have been dealt with. If temperature rises, this will not only lead to a higher sea level, but it will also mean that the whole climate will have changed. It is possible that during the winter precipitation will increase. This would not only affect the drainage system, it would also mean higher extreme river discharges of Rhine and Meuse, so again higher dikes (along the river) would be necessary. In summertime, evaporation could be higher and precipitation could be weaker. So shortage of fresh water during a dry summer or even during a normal summer could be greater, with all its consequences.

Another important aspect is the supply of drinking water, already mentioned when discussing saltwater intrusion. A major part of our drinking water supply comes from the fresh water lenses under the dunes. They are surrounded by saline water. To make it possible to subtract great amounts of water from these lenses, infiltration with fresh (river) water is currently necessary, because the precipitation surplus of the dunes is not enough. These lenses are used as a storage basin for drinking water. Moreover, the river water is filtered in this way, so it improves in quality. A sea level rise may distort the equilibrium of these lenses, especially where the dunes are rather small. Possible solutions would be: more infiltration of fresh water and extraction of less drinking water.

Ecological Impact - Distinguishing four areas where the impact of sea level rise is concentrated and taking a sea level rise of 1 m within 100 years, we get the following ecological impacts:

The total dune area of the central coast will diminish, and a strip of a width between 80 and 150 m will disappear. Considering a strengthening of the dunes (where needed) with sand depletion and not by dikes, the consequences for the ecology in this area will not be severe. So we assume that beach and coastal slope will be maintained, natural or artificial.

To the ecology of the Dutch estuarine systems the total extent of the intertidal area of the Wadden Sea area is very important. For some bird species this is the only place where they can feed. A decrease of the intertidal area will certainly result in a smaller number of birds per species and will probably also result in a smaller number of species. In the case of the Wadden Sea, the assumed sea level rise will probably hardly affect the extent of the intertidal area, so impact on ecology will be small. As mentioned previously the residual transports in the Wadden Sea may change. This can affect the ecological system, although it is still very difficult to predict in what way. The impact of temperature rise on the ecology is a totally different question, which cannot be answered yet.

In the Eastern and Western Sheldt of the Delta area the intertidal area is likely to decrease. This has the negative consequences mentioned above. The species concerned are mainly migratory birds (e.g., stilts).

In the lower parts of the river systems in the Netherlands, mean and extreme water levels will rise as well. The river forelands which now only get flooded occasionally during the winter will get flooded more frequently or will be flooded most of the time. The peculiar ecology of these will change or disappear.

Economic Implications - With a 1-meter sea level rise over a century, present maintenance costs for coastal and river defense systems would approximately double. The present cost of coastal maintenance amounts to about 60 million guilders per year. The strengthening of dikes, dunes, beaches, and shore faces over the next century would cost about 6 billion

guilders (3 billion U.S. dollars). Interestingly, the possibility of a storm surge barrier in the Rotterdam Waterway is currently under study as an alternative to the very high cost of strengthening the dikes along the lower river branches. The prospect of an increasing rate of sea level rise makes this alternative all the more attractive.

Although the structural adjustments required would be more complex than for the coastal defense, the amount of money involved is less. An investment of 3 billion guilders would be needed for pumps and changes in the infrastructure. The annual input costs (fuel) will increase by about 10 million guilders per year. Furthermore, harbours, locks, bridges, etc. would have to be adapted to the new situation; these costs would be roughly 1 billion guilders.

The total costs of adaptation to a 1-meter sea level rise over one century can be estimated at about 10 billion guilders (5 billion U.S. dollars). For comparison, the costs of the Delta Works (1958-1988) were about 14 billion guilders.

Conclusions

Mean sea level rise in the Netherlands has been between 15 and 20 cm per century. It is likely that sea level rise was not constant during the last 300 years and even negligible before 1850. During the past 20 years the Dutch coast has lost 5.3 million cubic meters of sand per year along the eroding parts and gained 3.4 million cubic meters per year along the accreting parts. As a whole, the Dutch coast is losing 1.9 million cubic meters per year.

A future acceleration of sea level rise will have great consequences for a low-lying country like the Netherlands. Due to a 1-meter sea level rise:

- The tidal motion in the North Sea and along the Dutch coast will hardly change.

- The storm surge heights will change insignificantly. The effect due to changes in storm frequency or severity of storms will be much more important, although at present it is unknown if frequency and severity will increase or decrease.

- The wave heights will change. To predict the changes in wave height it is necessary to know the morphological changes first.

- The dune coast line will retreat by 80 to 150 meters.

- The sedimentation in the Wadden Sea will increase and probably keep pace with this sea level rise, so mean water depth and intertidal area will probably remain constant.

- The salt intrusion via rivers and subsoil will increase. During summer the amount of fresh water needed for flushing of the polders will increase by about 10%.

- The water management of the Netherlands will have to be adapted. Large areas that now drain naturally by gravity will not do so in the future and large pumps will have to be installed.

- The ecological systems will be impacted. Intertidal area in the Eastern and Western Scheldt and valuable dune area along the coast will diminish. The ecological value of these area will be lost.

- The extra costs to raise our defenses against high water and to adapt our water management system will be about 10 billion guilders (5 billion U.S. dollars).

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